Geologic Map of the Estes Park 30′ x 60′ Quadrangle, North-Central Colorado

By James C. Cole and William A. Braddock

Pamphlet to accompany Scientific Investigations Map 3039

U.S. Department of the Interior
U.S. Geological Survey

CONVERSION FACTORS

Multiply	By	To obtain
centimeters (cm)	0.3937	inches (in.)
meters (m)	3.281	feet (ft)
kilometers (km)	0.6214	miles (mi)

To convert Celsius (°C) to Fahrenheit (°F), use formula (°C x 1.8) + 32

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By James C. Cole and William A. Braddock[1]

Abstract

The rocks and landforms of the Estes Park 30′ × 60′ quadrangle display an exceptionally complete record of geologic history in the northern Front Range of Colorado. The Proterozoic basement rocks exposed in the core of the range preserve evidence of Paleoproterozoic marine sedimentation, volcanism, and regional soft-sediment deformation, followed by regional folding and gradational metamorphism. The metasedimentary rocks of the Estes Park quadrangle are distinct within northern Colorado for preserving the complete metamorphic zonation from low-grade chlorite-muscovite phyllites, through middle greenschist-grade rocks with sequential aluminous porphyroblasts, to partially melted gneisses that contain high-grade cordierite and garnet in the non-melted residues. Regional and textural evidence shows that the widespread metamorphism was essentially concurrent with intrusion of the Boulder Creek Granodiorite and related magmas and with the peak of deformation in the partially melted high-grade rocks. The metamorphic thermal pulse arrived later following the peak of deformation in the physically higher, cooler, low-grade terrane.

Mesoproterozoic time was marked by intrusion of biotite granite in the Longs Peak-St Vrain batholith, a complex, irregular body that occupies nearly half of the core of the Front Range in this quadrangle. The magma was dry and viscous as it invaded the metamorphic rocks and caused wholesale plastic folding of the wall rock structure. Steep metamorphic foliation that resulted from the Paleoproterozoic deformations was bowed upward and re-oriented into flat-lying attitudes as the crystal-rich magma rose buoyantly and spread out in the middle crust. Magma invaded the schists and gneisses along weak foliation planes and produced a characteristic sill-upon-sill intrusive fabric, particularly in the higher parts of the batholith. Broad, open arches and swales that are defined by the flow-aligned feldspar foliation of the granite, as well as by compositional banding in the intruded and included metamorphic rocks, formed late during batholith emplacement due to rising, buoyant magma and sinking, dense wall rocks. The Longs Peak-St Vrain batholith was intruded into crust that was structurally neutral or moderately extending in an east-northeast direction. A broad zone of mylonite, the Moose Mountain shear zone, formed within the batholith during the final stages of consolidation as a result of differential buoyancy between the magma and dense wall rock, not as a result of regional tectonic deformation.

Direct evidence of the late Paleozoic Ancestral Rocky Mountains uplift is not present in this quadrangle, but the erosion of those highlands is recorded by Pennsylvanian and younger strata that unconformably overlie the Proterozoic basement rocks. The Phanerozoic sediments indicate a steady progression of fluvial, eolian, and lacustrine environments throughout most of the Mesozoic Era during times of relatively slow sediment accumulation. Early Cretaceous time was marked by incursion of the Western Interior seaway, a shallow-water marine embayment that persisted throughout the latter part of the Mesozoic Era. Cretaceous strata consist of abundant shale interbedded with relatively minor fine sandstone and limestone. Sedimentation rates increased markedly in the latter part of this period during downwarping related to distant crustal loading by thrusting along the western continental margin.

Mountain building resumed in this region during latest Cretaceous time with onset of the Laramide orogeny. This deformation produced regional basement uplifts bounded by steep faults. The eastern margin of the Front Range is marked by high-angle reverse faults and drape folds in the Phanerozoic strata, and by the deep syncline of the Denver Basin. The western range margin is locally marked by the low-angle Never Summer thrust that formed in Paleocene time following uplift of the eastern range margin. Mafic-alkalic and alkali-calcic magmas intruded the southern part of the quadrangle during the Laramide orogeny in the period from about 78 to 45 Ma; these intrusions are parts of an extensive northeast-trending province of igneous and hydrothermal activity known as the Colorado Mineral Belt and are host to base and precious metal deposits.

Post-Laramide time was marked by a prolonged period of weathering, erosion, and planation of the basement-rock surface, extending perhaps into late Oligocene or early Miocene time. Tuffaceous fluvial sediments of the

[1] Deceased, 2003

Troublesome Formation, interlayered with upper Oligocene through middle Miocene volcanic rocks, record this element of the geologic history on the western side of the Front Range. Intrusive activity in the Never Summer Mountains during middle Oligocene time led to eruption of andesitic, dacitic, and rhyolitic lavas and welded tuffs that blanketed the landscape. These rocks are collectively referred to as the Braddock Peak intrusive-volcanic complex.

Erosion on the eastern slope of the Front Range produced a broad, rolling surface surrounding residual highlands and east-trending fluvial channels filled with coarse, boulder gravel. Continued degradation of the mountainous highlands led to aggradation of a fluvial apron westward across the Laramide mountain front, probably equivalent to the Miocene Ogallala Formation on the High Plains. Low-gradient streams developed broad meandering paths across this aggraded apron. Renewed uplift of the Front Range during Pliocene time caused these meandering streams to incise rapidly and deeply into the Proterozoic basement rocks, producing the entrenched meanders that are characteristic of the major east-flowing streams west of the mountain front today.

Significant global cooling during the Pliocene set the stage for glaciation during the Quaternary. In the Estes Park quadrangle, the renewed mountain uplift accentuated the topographic relief across the Continental Divide and established the altitude necessary to trigger and accumulate abundant snow and persistent ice. Mountain glaciers advanced and retreated during at least three major glacial-interglacial cycles during the middle and late Pleistocene in this area. Erosion continues on the High Plains east of the mountain front, and progressive incision of the drainage is recorded by at least five major gravel-clad terrace and pediment surfaces along the major fluvial channels that connect to the South Platte River system.

The Estes Park quadrangle has diverse resources that have been exploited during the last 150 years. Early hard-rock and placer mining developed the hydrothermal deposits related to Laramide intrusions along the northern part of the Colorado Mineral Belt. Most of these deposits were vein fillings of gold and silver (native, arsenides, and tellurides) within fractures and breccias adjacent to the intrusions. The Boulder tungsten district produced iron tungstate (ferberite) and scheelite from hydrothermal veins and sericitic alteration zones. The Jamestown district produced fluorite, precious metals, and base-metal sulfides that accumulated in nested alteration zones around a younger, late Laramide syenite porphyry stock.

Coal was extensively mined from Upper Cretaceous Laramie Formation seams in the Boulder-Weld coal field between the 1850's and the 1970's. Oil and gas are produced from well fields in the eastern part of the quadrangle. The hydrocarbons were generated during depositional burial of source rocks (mostly Lower Cretaceous shale) in the Denver Basin. Oil and gas are produced from interbedded sand sheets within structural traps in the Laramide folds along the mountain front, as well as from stratigraphic truncations and pinch-outs.

Sand and gravel resources are widely developed in the stream-bed and fluvial-terrace deposits along all the major streams of the area. Dimension stone is quarried extensively from upper Paleozoic strata near the town of Lyons, and decorative rock is produced from the Dakota Group at many localities along the eastern mountain front. Tertiary dacite sills are quarried along the St Vrain Creek southwest of Lyons for road base, riprap, and landscaping dimension stone.

Water resources have been extensively developed throughout the region. Early settlers built systems of ditches and flumes to capture spring runoff from west of the Continental Divide and convey it to east-flowing streams for agricultural use on the Colorado Piedmont. Municipalities built high-altitude reservoirs in several drainages to store winter-spring snowpack and runoff for domestic supply and for agricultural use during the late summer. The most far-ranging water development project in the area is the Colorado-Big Thompson project that was constructed between 1938 and 1957 by the Bureau of Reclamation. It diverts as much as 220,000 acre-feet of water from the Colorado River drainage through a 13-mile tunnel beneath the Continental Divide to the Big Thompson River drainage. The total system diverts and stores water, generates power, and provides flood control, recreation, and regulated stream flow for much of northeastern Colorado.

The geology of the Estes Park quadrangle presents some risks and hazards for certain human uses of the land. The mountainous terrain involves inherent hazards related to rock fall and debris flow. Landslides have occurred in several settings in the area where slopes are moderate and bedrock contains significant silt and clay, especially in the Troublesome Formation, glacial till in the Colorado River valley, and in the Laramie Formation. The Dakota Group exposed in the foothills-hogback belt along the eastern mountain front has also generated distinctive block-glide landslides, in which entire sections of competent strata have detached along weak bedding planes and moved downslope. Many of these slides probably moved in the early Tertiary, but some may still be active today.

Flooding in mountain valleys has occurred in historic times in most of the major canyons of the Front Range. Summer thunderstorms statistically drop much of their precipitation between about 7,000 ft and 9,500 ft elevation, which coincides with the upper reach of the deeply incised canyons within the core of the range. The 1976 Big Thompson River flood demonstrated that these two conditions together can lead to intense, concentrated runoff, highly elevated stream power, and intense local damage along the narrow canyon floors.

Introduction

This report describes the geology of the Estes Park quadrangle in north-central Colorado. The compilation summarizes and integrates geologic mapping and various topical studies carried out between 1960 and 1990 under the

U.S. Geological Survey (USGS) study of the northeastern Front Range (Braddock and Cole, 1979; Hutchinson and Braddock, 1987), investigations of geology and land-use in the 1970's under the USGS Front Range Urban Corridor project (Trimble, 1975; Colton, 1978; Colton and Anderson, 1977; Hansen and Crosby, 1982), geologic mapping of Rocky Mountain National Park (Braddock and Cole, 1990), and detailed geologic mapping in northern areas of the Colorado Mineral Belt (Wrucke and Wilson, 1967; Gable and Madole, 1976; Gable, 1980a; Pearson, 1980). This publication also draws on topical studies in geology, hydrology, and geomorphology by numerous investigators from Federal, State, and local agencies, as well as university research published through the years.

The purpose of this compilation is to summarize the geologic framework of a significant portion of the Colorado mountain terrain in support of updated mineral-resource assessments, particularly in lands administered by the U.S. Forest Service.

Geography and Geomorphology

The Estes Park 30′ × 60′ quadrangle encompasses an area of 1,955 square miles, extends 57 miles east-west and 35 miles north-south, and includes parts of Boulder, Broomfield, Grand, Jackson, and Larimer Counties (fig. 1).

Approximately one-third of the land is Roosevelt, Routt, and Arapahoe National Forests, another one-third is Rocky Mountain National Park (RMNP), and a few thousand acres in the northwest corner lie within the Colorado State Forest. Wilderness areas are designated in some of the more remote, high alpine areas (both Forest Service and National Park lands), including the Never Summer, Neota, Indian Peaks, and Comanche Peak Wilderness areas (Pearson and others, 1981). A large area surrounding Lake Granby is designated the Arapahoe National Recreation Area.

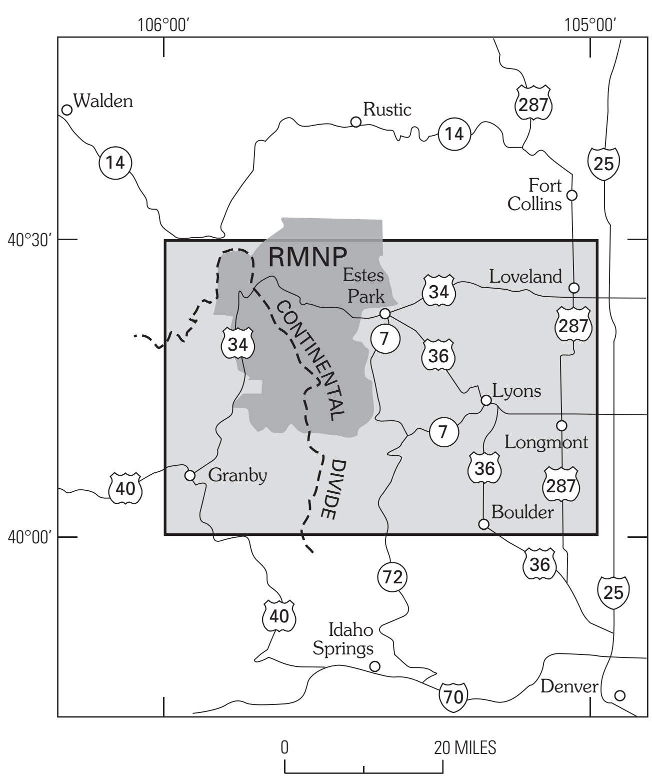


Figure 1. Map showing location of the Estes Park quadrangle and Rocky Mountain National Park (RMNP).

The southwestern corner of the quadrangle surrounds the mountain town of Granby at the confluence of the Colorado River and the Fraser River and is a mixture of private lands and public lands administered by the Bureau of Land Management (BLM).

Active and inactive mining claims cover significant acreage in and around the historic mining towns of Ward, Gold Hill, Wallstreet, Sunshine, and Jamestown.

The eastern one-quarter of the Estes Park quadrangle lies on the Colorado Piedmont (fig. 2) below about 6,000 ft altitude and includes the developed city centers of Boulder, Longmont, and Loveland, as well as interconnected suburban areas and significant farmland. The combined population of the Piedmont region is approximately 300,000 (2000 U.S. census) and contrasts significantly with the sparsely populated areas in the mountainous western regions.

Principal vehicle access in the area is provided by U.S. Highway 287 that passes north-south through the main cities in the east; U.S. Highway 34 that passes westward from Loveland, up the canyon of the Big Thompson River to Estes Park, and then over the Continental Divide through Rocky Mountain National Park (the highest paved-through highway in the nation) to Granby; Colorado State Highways 7 and 36 that connect mountain towns and the Piedmont; and U.S. Highway 40 through the Granby valley (fig. 1).

The Estes Park quadrangle spans the Front Range uplift of the Southern Rocky Mountains (Eaton, 1986) and the westernmost part of the Colorado Piedmont physiographic province to the east (Fenneman, 1931). The Continental Divide traces north to south across the quadrangle along the glaciated crest of the Front Range where numerous mountain summits exceed 12,000 ft elevation. Average timberline in this region is about 11,000 ft, and nearly 15 percent of the land surface of the quadrangle lies in this non-forested climatic zone of tundra.

The general landforms of the Front Range can be described in terms of five overlapping elevation-bounded zones within the quadrangle (fig. 2). The easternmost zone consists of the gently inclined slopes and broad fluvial valleys of the Colorado Piedmont. The eastern margin of the Front Range is marked by a foothill zone consisting of hogback ridges formed by upturned sedimentary strata of contrasting hardness. The foothill zone is narrow and linear in the southern quarter of the quadrangle; farther north, the range front is segmented by northwest-trending (fault-controlled) valleys that break up the hogback ridges into en echelon segments that step progressively eastward. The middle zone of the Front Range uplift (elevations between about 7,500 ft and 9,000 ft) is characterized by gently rolling upland block of Proterozoic crystalline rocks (Steven and others, 1997) where summits decline gently eastward but canyons are steep, narrow, and deeply incised. This rolling upland is interpreted by many to be a relic of widespread erosion in the middle Tertiary (Wahlstrom, 1947; Epis and Chapin, 1975; Scott and Taylor, 1986; Steven and others, 1997). The highest zone is the glaciated highland terrain above 9,000

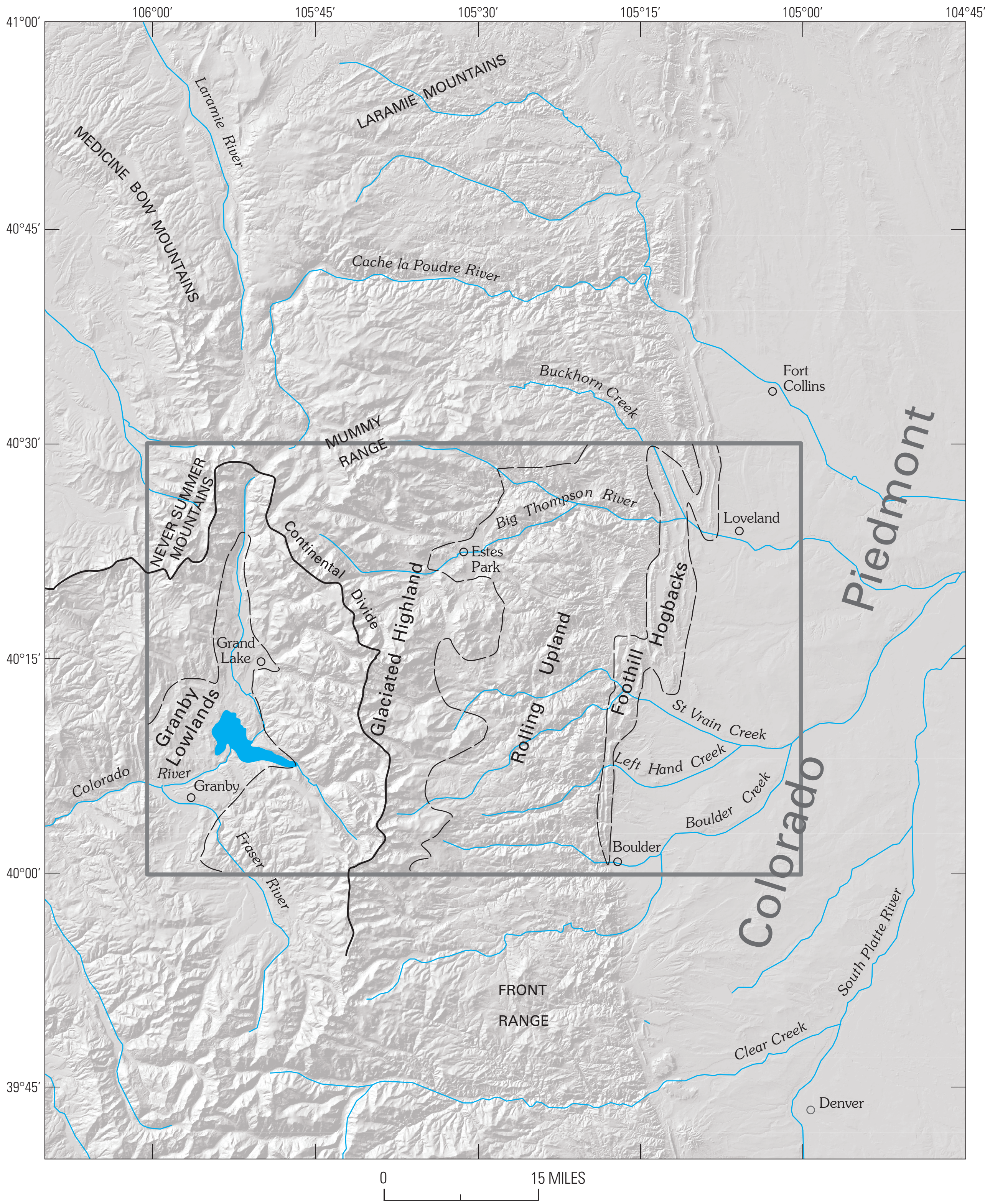


Figure 2. Map showing physiographic zones of the Estes Park quadrangle and adjoining areas.

ft elevation that is markedly steeper than the incised upland zone to the east; drainage reflects glacial processes of the Quaternary Period and the widespread glacial deposits formed in the valleys. The westernmost zone of the Front Range is the steep, western slope that declines rapidly to about 8,000 ft elevation in the fault-controlled north-south valley of the upper Colorado River, and the Tertiary sedimentary-basin lowlands surrounding Granby.

Several major rivers have headwaters along the Continental Divide within the Estes Park quadrangle. The westward-flowing Colorado River heads at La Poudre Pass at the north end of the Kawuneechee Valley, between the Never Summer Mountains on the west and the Front Range on the east. About four miles to the south at Milner Pass, drainage toward the northeast forms the headwaters of the Cache la Poudre River, and drainage toward the southeast forms the headwaters of the Big Thompson River (fig. 2). These two rivers traverse the Front Range and join the South Platte River east of Loveland near Greeley, Colorado. The southern part of the Front Range east of the Continental Divide in the Estes Park quadrangle drains into the St Vrain Creek through Lyons, or into Boulder Creek through Boulder, and then farther east to the South Platte River (Crosby, 1978).

Most of the eastern slope of the Front Range is rolling, forested upland. Local broad valleys and grass-covered flats are conspicuous variations to the more typical terrain, and are referred to as parks (Crosby, 1978). Many of these landforms were preferentially occupied by early settlers because slopes are gentler, bedrock is more deeply weathered, and farming and ranching were easier. These locations are designated by local place names: Estes Park, Meeker Park, Allens Park, Hermit Park, Rattlesnake Park, and Big Elk Park, to name a few. These landforms typically lie at altitudes between about 7,500 ft and 8,500 ft, where the steeper high-altitude terrain of the glaciated crest of the Front Range merges with the western margin of the uplifted, incised rolling upland, and where the through-flowing drainage has its gentlest gradient.

The foothills-hogback belt that marks the eastern margin of the Front Range uplift is a product of erosion during the last 5 million years (Eaton, 1987; Steven and others, 1997), not the result of recent mountain uplift. All drainage of the South Platte River system (fig. 2) has been incising more-or-less continuously during that interval. The major streams east of the mountain front are bordered by relict river terraces that record several periods of stillstand and local aggradation of floodplains. These deposits have been correlated with 5 principal periods of floodplain and alluvial-fan aggradation during the last 2 million years and they are linked to cyclic changes in discharge related to cyclic climate change and continental glaciation (Madole and others, 1998).

The U.S. Board of Geographic Names acted in 2008 to designate a previously unnamed summit (11,972 ft elevation) in the northwestern Never Summer Mountains as Braddock Peak, in recognition of W.A. (Bill) Braddock's lifelong achievements in geologic mapping, interpretation of Rocky Mountain geology, and teaching. The feature has been added to the base map for the Estes Park quadrangle in this publication. It is the namesake for the Braddock Peak intrusive-volcanic complex of Oligocene age.

Compilation Sources and Methods

Geology in most of the Estes Park 30′ × 60′ quadrangle was mapped at 1:24,000 scale under several USGS projects between the late 1960's and the 1990's (see map). Proterozoic rocks in the core of the Front Range were mapped by W.A. Braddock and numerous graduate students from the University of Colorado-Boulder through the Northeastern Front Range Project and published at 1:24,000 scale. Braddock compiled the geology of Rocky Mountain National Park at 1:50,000 scale from his and graduate students' mapping at larger scales (Braddock and Cole, 1990), and prepared the first digital geologic map database of the same area in the mid-1990's for use by National Park Service staff at RMNP.

The southern tier of 1:24,000-scale quadrangles was previously included in a 1:100,000-scale compilation of the Proterozoic crystalline rocks of the central Front Range (Gable, 2000). The present compilation was based on the same source maps and differs only slightly from Gable (2000), chiefly by simplification of the biotite-gneiss units.

The eastern three-eighths of the Estes Park quadrangle, along with additional geology to the north and east, was previously compiled at 1:100,000 scale by Colton (1978). Numerous areas, particularly within the Precambrian rocks of the Front Range and the complex hogback belt, were incorrectly labeled and portrayed on the 1978 compilation, and some fault relationships were omitted. For these reasons, the Colton (1978) compilation was not used as a primary source for the present work, except for the geology of portions of the eastern plains where most of the units are Quaternary surficial deposits.

Bedrock units east of the hogback belt on the Colorado Piedmont are largely covered by various surficial geologic units. Comprehensive regional study of the underlying Cretaceous marine deposits was completed by Scott and Cobban (1965, 1986) and incorporated in later geologic maps. Surficial geology was systematically mapped by Trimble (1975), Colton (1978), Colton and Anderson (1978), and by Madole and others (1998).

Braddock expanded the digital geologic map database for RMNP to include all of the Estes Park quadrangle. This work was completed in 1999 using commercial drafting software. The current digital geologic map database published with this report was converted to ArcInfo format by Ted Brandt (USGS, Denver) and modified slightly by Cole to incorporate additional and new information.

Acknowledgments

This geologic map compilation results from the singular dedication and persistence of W.A. (Bill) Braddock, involving hundreds of hours of meticulous digitizing over a period of several years. The work reflects his professional character and his devotion to careful, consistent geologic field mapping throughout his career of more than four decades. His geologic legacy is expressed not only by high-quality geologic map products like this, but also by the scores of geology students who learned field methods through his academic and professional leadership. He passed away in 2003 after completing the digital map compilation but before work was initiated on the descriptive and interpretive pamphlet. These latter items were completed by J.C. Cole.

Constructive comments and suggestions were incorporated based on technical reviews by Michael O'Neill, Ralph Shroba, and Cal Ruleman (USGS). The authors acknowledge and appreciate the thoroughness and thoughtfulness of their efforts. The final map and report are improved and refined as a result.

Ted Brandt converted the original digitized geologic map from AutoCAD files to ArcInfo coverages, audited and corrected quality-control issues, and constructed the digital databases for this project. His contributions were essential to completing this product.

Description of Map Units

[Map unit descriptions are summarized from descriptions in source maps and modified to reflect general characteristics of the unit across the Estes Park quadrangle. Color terms are descriptive of fresh rock/sediment material. Grain sizes of sediment particles are defined using the standard Wentworth size scale]

Alluvial Units

Qa **Alluvium (Holocene and upper Pleistocene)**—Stream-deposited sand, silt, and gravel underlying floodplains and adjacent low terraces along courses of major streams east of foothills-hogback belt of Front Range, and materials beneath floodplains in till-bounded glaciated valleys in mountains. Upper Pleistocene deposits are typically light gray and pale brown (clay and silt coatings on clasts); Holocene deposits are typically light pinkish gray due to abundance of detrital feldspar and general absence of clay-silt coatings (Madole and others, 1998). Locally includes low gravel-terrace remnants and sheetwash

Qva **Mountain valley alluvium (Holocene and upper Pleistocene)**—Gravel, sand, and silt along courses of major and intermittent streams in bedrock-cut valleys, in alluvial fans along montane valley margins, and in outwash deposits downstream from terminal moraines of Pinedale and Bull Lake age. Unit underlies narrow, high-altitude valleys west of Continental Divide, and montane valleys of Big Thompson and Little Thompson Rivers and St Vrain and Boulder Creeks east of Divide

Qbr **Broadway Alluvium (upper Pleistocene)**—Well-sorted and well-stratified sand and gravel beneath terraces. Top of unit is typically 10–25 ft above modern floodplains. Deposition was broadly coeval with Pinedale glaciation at 30 ka to 12 ka (Madole, 1991)

Qlv **Louviers Alluvium (upper and middle Pleistocene)**—Reddish-brown, clast-supported pebble and cobble gravel. Top of unit is typically 20–40 ft above modern floodplains. Deposition was broadly coeval with Bull Lake glaciation at about 300 ka to 120 ka (Madole, 1991)

Qs **Slocum Alluvium (middle Pleistocene)**—Reddish-brown, clast-supported pebble and cobble and boulder gravel and grayish-brown to brown silty sand. Top of unit is typically 110–130 ft above modern floodplains (Madole, 1991). Slocum also forms prominent pediment surface north of Boulder

Qv **Verdos Alluvium (middle Pleistocene)**—Brown sand and gravel. Pebbles, cobbles, and boulders are weathered and partly decomposed. Terrace and pediment deposits lie about 170 ft above Little Thompson River. Verdos Alluvium locally contains Lava Creek B ash (639 ka; Madole, 1991; Lanphere and others, 2002)

Qrf **Rocky Flats Alluvium (lower Pleistocene)**—Reddish-brown sand and clast supported pebble and cobble gravel and brown silty and clayey sand. Preserved at Table Mountain north of Boulder and Table Top Mountain north of Hygiene (Madole and others, 1998), where upper surface is typically 260–300 ft above modern floodplains

Qprf **Pre-Rocky Flats alluvium (lower Pleistocene)**—Brown to white carbonate-cemented gravel and sand; only preserved on Haystack Mountain and Gun Barrel Hill northeast of Boulder. Top of gravel (eroded) is more than 370 ft above modern floodplains

Qgy **Young gravel deposits at Granby Mesa (upper Pleistocene)**—Gravel and sand deposits underlying fluvial terraces along Colorado and Fraser Rivers adjacent to Granby Mesa; tops of terraces approximately 60 ft above modern floodplains. Terrace deposits probably formed during Pinedale glaciation (Meierding, 1977; R. Shroba, written commun., 2006)

Qgm **Intermediate gravel deposits at Granby Mesa (upper? and middle? Pleistocene)**—Gravel and sand deposits underlying fluvial terraces along Colorado and Fraser Rivers adjacent to Granby Mesa; tops of terraces approximately 140 ft above modern floodplains. Terrace deposits probably formed during Bull Lake glaciation (Meierding, 1977; R. Shroba, written commun., 2006)

Qgo **Old gravel deposits at Granby Mesa (middle? and lower? Pleistocene)**—Gravel and sand deposits underlying fluvial terraces remnants on Granby Mesa; tops of terrace remnants as high as 220 ft above modern floodplains. Terrace deposits probably formed prior to Bull Lake glaciation (Meierding, 1977; R. Shroba, written commun., 2006)

Qfo **Old alluvial-fan deposits (lower Pleistocene)**—Poorly exposed, indurated sand and gravel deposits on east flank of Granby Mesa that underlie an eroded alluvial fan landform; unconformably overlie tilted Mesozoic bedrock units; flanked westward by old gravel deposits of Granby Mesa (Qgo). Thickness unknown

Qpy **Younger piedmont-slope alluvium (upper and middle Pleistocene)**—Chiefly brown to reddish-brown sand, silty sand, and matrix-supported gravel derived from upturned sedimentary rocks

Qd **Diamicton (upper? Pleistocene)**—Bouldery gravel and sand deposits on the floor of Tahosa Valley that lie beyond known limits of glacial ice; poorly exposed, but surface displays boulders as large as 4 ft. Formerly interpreted to be pre-Bull Lake till (Richmond, 1960), but deposits are probably younger because they are low in landscape near modern drainage. Form of diamicton deposit near Meeker Campground suggests relict alluvial fan (Braddock and Cole, 1990)

Ng **Fluvial gravel (Pliocene? and Miocene)**—Unsorted, unstratified deposits of boulders (as large as ten ft), cobbles, and pebbles in a dominantly coarse sandy matrix with some silt; in Niwot Ridge area of Ward quadrangle, deposits are higher on landscape than highest glacial deposits of Bull Lake age, are much thicker (as great as 100 ft) and more voluminous than typical till, and are unlikely to have formed by glacial processes (Gable and Madole, 1976). Madole (1982) concluded these deposits formed in ancient valley systems that were subsequently disrupted by Neogene uplift of the Front Range. These high-level fluvial-gravel deposits also mark paleo-canyons of the Big Thompson River (north of Palisade Mountain) and the early Colorado River drainage (Gravel Mountain and north of Apiatan Mountain).

Glacial Units

Qr **Rock glacier deposit (Holocene and upper Pleistocene)**—Lobate mass of rock rubble, silty sand, and ice that is veneered by angular blocks and boulders in alpine cirques and valleys above 10,000 ft elevation. Surface typically marked by conspicuous arcuate ridges and a steep lobate front (Gable and Madole, 1976)

Qo **Organic-rich sediment (Holocene and upper Pleistocene)**—Bog and marsh deposits in formerly glaciated terrain, locally overlain by glacial-outwash sediments

Qtp **Till of Pinedale age (upper Pleistocene)**—Subangular to subrounded boulders, cobbles, and pebbles in a silty sand matrix. Prominent, sharp-crested moraines are numerous, most clasts slightly weathered, soils thin. Contains small bodies of Holocene till in cirque basins. Tills in Arapahoe Creek drainage east of Lake Granby are inferred to be Pinedale age, but may include local remnants of older glacial deposits; detailed work has not been conducted in this drainage. Pinedale glaciation occurred between about 30 ka and 12 ka in this region (Madole, 1991)

Qtb **Till of Bull Lake age (upper and middle Pleistocene)**—Subangular to subrounded boulders, cobbles, and pebbles in a silty sand matrix. Morainal form subdued, clasts variably weathered, and soils well developed. Bull Lake glaciation occurred between about 300 ka and 120 ka in this region (Madole, 1991)

Qtpb **Till of pre-Bull Lake age (middle and lower? Pleistocene)**—Weathered boulders and cobbles in a silty sand matrix. Deposits lack morainal form and have thick reddish-brown clayey soil containing decomposed granitic clasts. Pre-Bull Lake deposits are identified in Colorado River valley near Shadow Mountain Lake and Granby Dam (Meierding, 1977; Braddock and Cole, 1990)

Other Surficial Units

f **Manmade fill (Holocene)**—Engineered deposits in earthen dams in numerous locations throughout quadrangle. Includes large area of transported rock and soil surrounding a limestone quarry and cement manufacturing facility east of Lyons

Qc **Colluvium (Holocene and upper Pleistocene)**—Deposits of rock debris ranging in grain size from silt to boulders, formed by a variety of mass-wasting processes. Unit chiefly includes small-volume deposits in piedmont and marginal to montane valleys, as well as extensive areas near and above timberline where freeze-thaw action has aided mass wasting and downslope movement

Qac **Alluvium and colluvium (Holocene and upper Pleistocene)**—Thin deposits of silt, sand, and gravel in strike valleys within hogback physiographic zone of eastern foothills. Most debris is derived from resistant rocks in flanking strike ridges; transported downvalley by episodic, concentrated streamflow

Qlo **Loess (Holocene and upper? Pleistocene)**—Pale-brown sandy to clayey silt deposited by wind. Unit chiefly confined to low-relief areas east of Front Range where it forms a thin blanketing deposit, mostly 3–10 ft thick. Locally may include minor deposits of middle Pleistocene age

Qls **Landslide deposits (Holocene and upper Pleistocene)**—Includes slumps, earth flows, rock-avalanche deposits, debris flows, and block-glide landslides. Stratigraphically controlled block-glide landslides are common on dip slopes of Dakota Group east of Front Range (Braddock, 1978). Landslides also common in glacial deposits on slopes of Kawuneeche Valley north of Lake Granby, and in Troublesome Formation near Granby. Several landslide deposits of Proterozoic metamorphic rock on and near Jackstraw Mountain appear to have moved on weak foliation planes inclined downslope (Braddock and Cole, 1990). Prominent landslide on northwest slope of Porphyry Peaks involves hydrothermally altered Oligocene subvolcanic intrusive rocks; jumbled blocks and irregular landform suggests later Pleistocene or younger age

Qta **Talus deposits (Holocene and upper Pleistocene)**—Angular blocks of rock on steep slopes below cliffs. Unit is chiefly shown in alpine and subalpine areas along crest of Front Range and in Never Summer Mountains

Post-Laramide Sedimentary Rocks

N₱t **Troublesome Formation (lower Miocene and upper Oligocene)**—Gray and orange-gray, tuffaceous mudstone and sandstone, volcanic ash beds, and minor clayey limestone and conglomerate; locally interbedded with basalt and trachyandesite (₱ba). Unit is widely exposed in southwestern corner of quadrangle; base is a significant unconformity with as much as 650 ft of local relief on older rocks (Braddock and Cole, 1990; Schroeder, 1995a). Lower part of Troublesome Formation is locally interlayered with Oligocene basalt (26.4 ± 0.9 Ma) and latite breccia (28.4 ± 2.3 Ma; Izett, 1974)

₱s **Sedimentary rocks, undivided (Oligocene and Eocene?)**—Small exposures of sedimentary rocks that interfinger with or underlie Oligocene volcanic deposits in northern part of Kawuneechee Valley

Post-Laramide Intrusive Rocks

N₱i **Intrusive rocks, undivided (Miocene to Oligocene)**—Dikes, sills, and small plugs; typically very fine grained and altered; compositions range from basalt and andesite to rhyolite. Intrusions are chiefly located in Never Summer Mountains and westward in Illinois River drainage

₱gc **Granite and rhyolite porphyry of Mount Cumulus stock (late Oligocene)**—Massive, tan to orange-red, leucocratic intrusive rocks that grade from rhyolite porphyry (south and west) to coarsely crystalline granite (north and east); intruded at 28.2 ± 0.7 Ma (whole-rock K-Ar; Marvin and others, 1974)

₱rp **Rhyolite porphyry (late Oligocene)**—White, tan, or light-reddish-brown porphyritic rock containing 20–40 percent phenocrysts of quartz, sanidine, oligoclase, biotite, and brown hornblende in an aphanitic matrix

₱bi **Basalt intrusion (late Oligocene)**—Dark-gray to black, dense rock intruded into lower part of Troublesome Formation (N₱t) with interbedded basalt and trachyandesite flows (may be source of latter). Single body located about 2 mi west of northwestern Lake Granby shore

₱gdr **Granodiorite and monzonite of Mount Richthofen stock (late Oligocene)**—Light- to medium-gray equigranular to slightly porphyritic granodiorite, monzonite, and quartz monzonite; compositional facies are all intergradational and all contain hornblende, biotite, and minor augite; intruded at 29.7 ± 3 Ma (whole-rock K-Ar; Marvin and others, 1974)

₱ti **Intrusive tuff-breccia (late? Oligocene)**—Yellowish-gray to grayish-yellow-green, poorly sorted, weakly bedded tuff-breccia consisting of volcanic ash and lapilli, comminuted Proterozoic rocks, sparse siltstone and sandstone of possible Mesozoic age, and Oligocene igneous rocks (andesite porphyry and quartz latite). Forms a singular pipe-like body northwest of Bowen Mountain within Proterozoic rocks (Metzger, 1974); planar fabric defined by clast sorting dips gently inward

₱qla **Quartz latite of Apiatan Mountain (late Oligocene)**—Gray to brownish-gray porphyry containing about 25 percent phenocrysts of oligoclase-andesine (as long as 3 cm), sanidine, and minor biotite and clinopyroxene in a microcrystalline groundmass rich in potassium feldspar; intruded at 28.4 ± 2.3 Ma (whole-rock K-Ar; Marvin and others, 1974)

₱qlx **Quartz latite breccia (upper Oligocene)**—Brown to reddish-brown monolithologic breccia; locally grades into quartz latite (₱qla); breccia is probably mostly extrusive, and interbedded with lower part of Troublesome Formation (N₱t)

Post-Laramide Volcanic Rocks

₱r **Rhyolite (upper Oligocene)**—White, light-gray, pale-yellowish-brown, moderate-brown, or pale-red-purple, flow-laminated rhyolite with 5–10 percent phenocrysts of sanidine, quartz, and minor biotite; flow laminations are typically contorted; probably coeval with granite and rhyolite porphyry of Mount Cumulus stock (₱gc). Rhyolite flows generally overlie rhyolite welded tuff (₱rw) north of Thunder Mountain; similar flow rock overlies Troublesome Formation (N₱t) east of the Kawuneechee Valley opposite mouth of Baker Gulch

₱rw **Rhyolite welded tuff (upper Oligocene)**—Strongly welded, weakly layered gray, black, or brown crystal-rich tuff distinguished by brown, euhedral smoky quartz phenocrysts; probably coeval with granite and rhyolite porphyry of Mount Cumulus stock (₱gc). Erupted at 27.7 ± 0.1 Ma (sanidine ^{40}Ar/^{39}Ar; Knox, 2005) from the Braddock Peak intrusive-volcanic complex

₱v **Volcanic rocks, undivided (upper Oligocene)**—Complexly interlayered volcanic-flow rocks and volcanic-rich sedimentary rock in Never Summer Mountains and in Specimen Creek area north of Trail Ridge Road. Major rocks include rhyolite, rhyolite-welded tuff, heterolithic debris-flow deposits, and lesser andesite and volcaniclastic sandstone

₱vx **Volcanic breccia (upper Oligocene)**—Massive to layered breccia containing fragments of older volcanic rocks, Phanerozoic sedimentary rock (as large as 12 in) and blocks of Proterozoic crystalline rock (as large as 35 ft; Lee, 1917; Braddock and Cole, 1990), typically interbedded with thick rhyolite welded tuff. Character and spatial distribution suggest these deposits formed during eruption of rhyolite welded tuff (₱rw) in Never Summer Mountains

₱ba **Basalt and trachyandesite (upper Oligocene)**—Dark-gray, very fine grained rock that weathers to grayish brown, grayish purple, or moderate red. Interlayered with the Troublesome Formation (N₱t) in southwest part of map area, and with other volcanic rocks in northwest part; erupted at 26.4 ± 0.9 Ma (whole-rock K-Ar; Izett, 1974)

₱ap **Andesite porphyry (upper? Oligocene)**—Dark-gray to black, massive rock that contains abundant yellowish-white andesine phenocrysts and less common orthopyroxene, clinopyroxene, biotite, and hornblende in a partly glassy to cryptocrystalline matrix

Laramide Sedimentary and Volcanic Rocks

₱c **Coalmont Formation (Eocene and Paleocene)**—Fluvial sedimentary unit within North Park basin consisting of a lower unit of gray and green mudstone and an upper unit of fine- to coarse-grained tan arkose with lenses of coarse conglomerate; thickness exceeds 500 ft in Jack Creek drainage (top truncated by Never Summer thrust fault). Formation is partly the lateral equivalent of Middle Park Formation (₱Km) deposited in Middle Park basin; arbitrary boundary between units is Continental Divide south of Illinois River. Coalmont Formation lies on volcaniclastic sedimentary rocks containing common clasts of andesite, which might be

Windy Gap Volcanic Member of Middle Park Formation (O'Neill, 1981) or "volcanic pebble member" (lower part) of Coalmont Formation as recognized by Kinney (1970b) in area immediately west

Middle Park Formation (Paleocene and Upper Cretaceous?)

₱Km **Upper part (Paleocene and Upper Cretaceous?)**—Interbedded, light- to medium-brown, tan or gray, volcanic and (or) arkosic sandstone and siltstone; volcanic and (or) granitic pebble to boulder conglomerate; and red, green, and brown mudstone. Thickness exceeds 5,000 ft in region west of this quadrangle in Middle Park basin (Izett, 1968)

Kmw **Windy Gap Volcanic Member (Upper Cretaceous?)**—Brown or greenish-brown, gray, or purple, massive volcanic breccia; interbedded volcanic conglomerate and sandstone; and rare, probable flows of latite to trachyte porphyry. Age is inferred from fossil leaves in overlying beds of upper part of Middle Park Formation (₱Km) (Izett, 1968). Thickness about 300 ft at western quadrangle boundary west of Granby. Beds of volcaniclastic conglomerate and sandstone north of Illinois River beneath Never Summer thrust lack volcanic breccia but were correlated with Windy Gap Member by O'Neill (1981); these andesite-clast-bearing beds may be lower Coalmont Formation (₱c) (Kinney, 1970b)

Laramide Intrusive Rocks

₱sy **Syenite and quartz syenite (Paleocene and Eocene)**—Light-gray, slightly pinkish-gray weathering, leucocratic porphyritic rock with a fine-grained, dense, granular groundmass. Syenite in Audubon-Albion stock may be Paleocene based on unpublished biotite age of 65 Ma ($^{40}Ar/^{39}Ar$; DeWitt, written commun., 2007). Syenite also forms numerous small plugs and irregular bodies near Jamestown and Sunset that were intruded at 56 and 45 Ma (Gable, 1980a; DeWitt, written commun., 2007). Quartz syenite grades to hornblende quartz monzonite in the Tuscarora stock (DeWitt, written commun., 2007)

₱ql **Quartz latite and rhyodacite porphyry (Paleocene)**—Gray-brown and greenish-gray porphyritic intrusive rocks with phenocrysts of altered pale-yellow feldspar and biotite or hornblende. Porphyries intruded as tabular sills into tilted sedimentary strata in foothills-hogback belt between Lyons and Boulder; ages range between 66 Ma and 64 Ma (Hoblitt and Larson, 1975; DeWitt, written commun., 2007). Similar rocks form common cross-cutting dikes in the Indian Peaks area (Pearson, 1980) and across the Colorado Mineral Belt from Jamestown to Arapaho Pass

Klp **Augite latite porphyry (Late Cretaceous)**—Medium- to dark-greenish-gray porphyry with phenocrysts of plagioclase and augite intruded into Cretaceous strata south of Granby. Probable age is 66.0 ± 4.2 Ma (fission-track; Schroeder, 1995a)

Kqm **Quartz monzonite (Late Cretaceous)**—Gray, equigranular rock that consists of feldspar, hornblende, biotite, and minor quartz; gradational to quartz syenite. Mapped in three small bodies in Jamestown-Ward area, including Tuscarora stock, Long Gulch stock, and an unnamed plug near northeast margin of Boulder Creek batholith

Ksyd **Syenodiorite (Late Cretaceous)**—Gray, medium-grained rock commonly with a salt-and-pepper appearance, but locally more mafic and darker. Syenodiorite forms central part of Audubon-Albion stock intruded at about 68 Ma (whole-rock K-Ar, Gable and Madole, 1976; $^{40}Ar/^{39}Ar$, DeWitt, written commun., 2007); locally grades to syenogabbro

Kmzd **Monzodiorite (Late Cretaceous)**—Gray, medium-grained, equigranular to slightly porphyritic rock consisting chiefly of feldspar and about 10–15 percent hornblende. Intruded into Jamestown stock between about 78 Ma and 72 Ma (Gable, 1980a, 1984; DeWitt, written commun., 2007); largely unmineralized. Intruded into Caribou stock at about 76 Ma (DeWitt, written commun., 2007); largely unmineralized

Kmzb **Monzogabbro (Late Cretaceous)**—Dark-gray, salt-and-pepper, fine- to medium-grained, slightly porphyritic rock showing a crude flow banding with abundant mafic clots; forms early cumulate masses in western part of Audubon-Albion stock. Age is about 70 Ma (DeWitt, written commun., 2007)

Pre-Cenozoic Sedimentary Rocks

Kl **Laramie Formation (Upper Cretaceous)**—Light-gray to light-yellowish-gray sandstone, sandy shale, claystone, shale, and several beds of economic coal in lower part. Laramie Formation was deposited in coastal plain and swamp environments, generally coeval with regressive Fox Hills Sandstone that was deposited along coastline (Roberts, 2005). Thickness ranges between 350 ft and 1,800 ft in the region

Kfh **Fox Hills Sandstone (Upper Cretaceous)**—Upper part of formation consists of crossbedded tan sandstone; grades downward into brown, fine-grained silty sandstone and gray shale. Fox Hills formed in strandline and delta-front environments during withdrawal of Late Cretaceous Western Interior seaway (Roberts, 2005); thickness variable because Fox Hills consists of several overlapping tabular sand bodies, but total thickness locally as great as 300 ft

Pierre Shale (Upper Cretaceous)—Formation has been subdivided into a number of members on east side of map, but is undivided in Never Summer Mountains where it is deformed. Units are well described by Scott and Cobban (1965; 1986); total thickness is as great as 6,200 ft in Denver Basin (Higley and Cox, 2005)

Kp **Pierre Shale, undivided**—Primarily dark-gray to black silty shale with some beds of sandstone. Adjacent to Mount Richthofen stock, Pierre Shale has been thermally metamorphosed to dense, hard, medium- to dark-gray hornfels and recrystallized sandstone (O'Neill, 1981). Elsewhere, subdivided as follows:

Kpt **Upper transition member**—Friable sandstone and soft shaley sandstone containing large calcareous concretions

Kpu **Upper shale member**—Gray, concretionary silty shale

Kpr **Richard Sandstone Member, Larimer Sandstone Member, Rocky Ridge Sandstone Member and two intervening unnamed shale members**—Light-brown or brownish-yellow, micaceous or glauconitic sandstone interlayered with gray shale

Kpm **Middle shale member**—Claystone and sandy siltstone

Kph **Hygiene Sandstone Member**—An upper hard, glauconitic, ridge-forming sandstone separated from a lower friable sandstone by a shale

Kpl **Lower shale member**—Mostly dark-olive-gray bentonitic shale

Kn **Niobrara Formation (Upper Cretaceous)**—In western map area, formation consists of light-gray-weathering, platy, calcareous shale; weathers buff near top; thin, light-gray biomicrite beds near middle; thin, light-gray micritic limestone at base. In eastern map area, formation consists of two members (combined thickness about 320 ft):

Smoky Hill Shale Member—Very fissile calcareous shale; dark gray on fresh surfaces, weathers to light gray plates; distinctive yellowish-brown-weathering micrite at top; thickness about 300 ft

Fort Hays Limestone Member—Light-gray, thick-bedded micrite; thickness about 20 ft

Kbm **Benton Shale (Upper Cretaceous) and Mowry Shale (Lower Cretaceous), undivided**—Combined unit of several shale, calcareous shale, and limestone units that lie above Dakota Group and below Niobrara Formation; exposure is typically poor. Benton Shale is locally subdivided (from top to bottom) into the olive-gray, silty-sandy Carlile Shale, dark-gray to olive-gray Greenhorn Limestone, and dark-gray Graneros Shale. The underlying Mowry Shale (locally absent) is siliceous and weathers light gray. Combined units are about 350 ft thick in Granby area, and about 530 ft thick along eastern Front Range margin

Kd **Dakota Group or Sandstone (Lower Cretaceous)**—In western map area, Dakota Sandstone consists of light-gray, fine-grained sandstone and interbedded shale in upper part, and chert-pebble conglomerate and sandstone in lower part; total thickness about 175–230 ft (Izett, 1968). In eastern map area, Dakota Group has generally been mapped as two formations:

South Platte Formation

First sandstone member—Gray, well-sorted, fine- to medium-grained sandstone; 40–100 ft thick in outcrop. Unit correlates with Muddy Sandstone (or informal "J" sandstone) in subsurface (Higley and Cox, 2005)

Middle shale member—Dark-gray carbonaceous shale and siltstone; 90 ft thick

Plainview Sandstone Member—Gray to light-brown, thinly bedded, fine-grained sandstone; about 30 ft thick

Lytle Formation—Gray to light-brown, coarse-grained to conglomeratic sandstone and blocky-weathering, varicolored, noncarbonaceous mudstone; about 120 ft thick. Base of Lytle Formation is a regional unconformity

Jm **Morrison Formation (Upper Jurassic)**—Green, red, yellow, and white blocky-weathering claystone, siltstone, and gray micrite, and gray, fine- to medium-grained sandstone; about 300 ft thick

Jms **Morrison Formation (Upper Jurassic) and Sundance Formation (Middle Jurassic), undivided**—Morrison Formation is similar to description above (Jm). Sundance Formation is buff, very fine grained sandstone and laminated siltstone; thickness about 125 ft west of Front Range

J₸mj **Morrison Formation (Upper Jurassic), Sundance Formation (Middle Jurassic), and Jelm Formation (Upper Triassic), undivided**—Morrison Formation is similar to description above (Jm). Pine Butte Member of the Sundance Formation is tabular-bedded, fine-grained, gray to white sandstone; it conformably overlies Canyon Springs Sandstone Member of Sundance which is pink, orange-pink, or reddish-brown, fine- to medium-grained, crossbedded, calcareous sandstone. Sundance unconformably overlies Red Draw Member of Jelm Formation, which is orange-pink, or reddish-brown, fine-grained, crossbedded, calcareous sandstone. Combined thickness ranges from about 350 ft (southeast) to 420 ft (northeast). Base of Jelm Formation is a regional unconformity

₸Pc **Chugwater Formation and older units (Lower Triassic and Upper Permian)**—Reddish-brown to orange-red shale, siltstone, and fine-grained sandstone; laminated to thin bedded. In northwestern map area, lower part contains light-gray to white finely laminated limestone (possible Forelle Limestone Member of Late Permian age). In southwestern map area, formation contains thin, lenticular limestone at base. Thickness ranges from 300 ft (southwest) to 830 ft (northwest). Entire unit unconformably overlies Proterozoic basement west of Front Range

₸Pl **Lykins Formation (Lower Triassic and Upper Permian)**—In the eastern foothills-hogback belt

Upper part (Lower Triassic and Upper Permian)—Maroon or reddish-brown siltstone and sandstone; thin limestone near base; about 370 ft thick

Forelle Limestone Member (Upper Permian)—Yellowish-brown and yellowish-gray, finely laminated, wavy-bedded limestone with stromatolite structures; about 30 ft thick

Lower part (Upper Permian)—Moderate-reddish-brown silty shale and minor limestone; about 100 ft thick

Pl **Lyons Sandstone (Lower Permian)**—Medium-orange, pink, or pinkish-gray, fine- to medium-grained, silica-cemented, well-sorted, cross-stratified, quartz sandstone; deposited in coastal-dune complex; about 60 ft thick. Base of Lyons Sandstone is a regional unconformity

Po **Owl Canyon Formation (Lower Permian)**—Red siltstone and fine-grained, red, ripple-laminated sandstone deposited in shallow-water marine estuary environment; thickness as great as 250 ft (north) but formation pinches out south of Little Thompson River

Pi **Ingleside Formation (Lower Permian)**—Pink to light-red, fine-grained quartzose sandstone. Commonly well cemented with quartz or calcite; thin to very thick bedded, locally crossbedded; deposited in a shallow-water marine-shelf environment; thickness varies from 50 ft (north) to about 200 ft (south)

PIPf **Fountain Formation (Lower Permian and Upper and Middle Pennsylvanian)**—Reddish-brown to purplish-gray, arkosic conglomerate; medium- to coarse-grained, feldspathic sandstone, trough crossbedded; dark-reddish-brown siltstone and shale; and minor thin limestone; deposited in alluvial fans and coastal-plain environments; about 1,100 ft thick

MzPz **Sedimentary and metamorphic rocks (Mesozoic and Paleozoic), undivided**—Rocks exposed in small fault blocks and slivers along trace of Never Summer thrust and as inclusions in granitic stock of Mount Cumulus (see O'Neill, 1981, for greater detail)

Paleozoic(?) Intrusive Rocks

Dk **Kimberlite (Devonian?)**—Dark-gray or dark-olive-green, altered, fragmental rock; contains pale-golden-yellow phlogopite, small grains of purplish-red pyrope garnet, ilmenite, and chlorite psuedomorphs of pyroxene(?); occurs as small dikes 1.3 mi east of Rams Horn Mountain, and talus blocks in Hayden Gorge. Early Devonian age determined on phlogopite from dikes (Smith, 1979). Hayden Gorge kimberlite may be similar or older, based on radiometric ages obtained elsewhere in the northern Front Range that indicate three periods of eruption at about 350 Ma, 570 Ma, and 620–640 Ma (Lester and others, 2001; Lester and Farmer, 2004)

Mesoproterozoic Intrusive Rocks

Proterozoic intrusive rocks in the Estes Park quadrangle were emplaced during two principal time periods, based on radiometric age determinations by several methods. Rocks formed during these major intrusive periods were recognized and formalized by Tweto (1987) as the Paleoproterozoic Routt Plutonic Suite (roughly 1,700 ± 25 Ma) and the Mesoproterozoic Berthoud Plutonic Suite (roughly 1,400 ± 25 Ma). Batholith-sized intrusions have long been designated with a geographic place-name that has served as a type locality for a named lithodemic unit (for example, Boulder Creek batholith consisting primarily of Boulder Creek Granodiorite; or Silver Plume batholith consisting primarily of Silver Plume Granite). Rocks of similar composition, texture, and age have been recognized in separate batholithic bodies at numerous locations in the Colorado basement, and some have been correlated (by these characteristics) with the named lithodemic units (see Tweto, 1987). Later isotopic studies have shown that a few of these correlations were not valid (Premo and others, 2007)

The nomenclature and symbology for Proterozoic intrusive rocks in this quadrangle (and adjoining areas) are based on age, composition, and a local intrusive-body name, rather than on inferred correlation to the type of a named lithodemic unit.

Map-unit symbol is composed as follows: intrusion age (X = Paleoproterozoic 1,700 ± 25 Ma; Y = Mesoproterozoic 1,400 ± 25 Ma); composition (g = granite; gd = granodiorite; d = diorite; qd = quartz diorite; gb = gabbro; j = trondhjemite); and intrusive body (B = Boulder Creek batholith; H = Granite of Hagues Peak; LP = Longs Peak-St Vrain batholith; T = Thompson Canyon intrusions).

Textural and mineralogic variants separately mapped within Longs Peak–St Vrain batholith are symbolized with an additional lower-case letter: s = sillimanite-bearing; x = intrusion breccia

Ygb **Gabbro of the Iron Dike**—Dark-gray to black ferrogabbro; weathers dark brown to orange brown with prominent limonite stains along joint surfaces; very fine grained along dike margins, and medium grained in centers of thicker dike segments (as wide as 50 ft). Intruded 1,316 ± 50 Ma (Rb-Sr isochron; Braddock and Peterman, 1989). Iron Dike forms a narrow north-northwest trending swarm that can be traced almost continuously from eastern margin of Front Range near Boulder, across Estes Park quadrangle, and beyond into Medicine Bow Range to near Colorado-Wyoming State line, over a distance of about 95 mi

Yga **Biotite-muscovite alkali-feldspar granite**—Gray to buff, fine- to medium-grained, xenomorphic granular granite with sparse accessory biotite and possible magmatic muscovite. Grades into pegmatite in a widespread intrusive complex near Drake; intruded at 1,361 ± 30 Ma (Rb-Sr isochron; Peterman and others, 1968)

Ya **Aplite**—Cream to buff, fine- to medium-grained, xenomorphic-granular rock with minor accessory magmatic sillimanite or garnet. Gradational with pegmatite west of Glen Haven

YgLP **Granite of Longs Peak batholith**—Light- to medium-gray, grayish-orange, orange-pink, or red-purple monzogranite to syenogranite that contains characteristic tabular microcline phenocrysts. Biotite is the principal dark mineral (5–12 percent), locally accompanied by magmatic sillimanite and (or) garnet plus minor accessory minerals. Equigranular matrix is locally fine, medium, or coarse grained and weakly foliated by aligned biotite grains. Granite intruded the Longs Peak-St Vrain batholith as a viscous magma that extensively deformed metamorphic wall-rocks and intruded as tabular sills in the center (highest preserved parts) of the batholith between rotated, flat-lying foliation planes in biotite gneiss (Cole, 1977; Braddock and Cole, 1979, 1990). Intruded at 1,420 ± 25 Ma (Rb-Sr isochron; Silver Plume Granite of Peterman and others, 1968). Secondary muscovite (non-oriented, porphyroblastic) is locally prominent, particularly along the northeastern margin of the batholith and in small satellite intrusions; formed by subsolidus hydration of sillimanite + potassium feldspar or breakdown of primary biotite to muscovite + magnetite.

YgLPx **Intrusion breccia**—Large irregular area (about 0.5 sq mi) east of Stones Peak consisting of intrusive matrix of biotite-rich granite of Longs Peak (YgLP) containing about 75 percent rounded to angular fragments of biotite gneiss, pegmatite, amphibolite, and medium-grained porphyritic granite; similar breccia occurs in small (unmapped) areas near Sky Pond and just north of Hidden Valley parking area. Probably formed during late-stage, explosive exsolution of water vapor during crystallization of granite of Longs Peak (Cole, 1977)

YgLPs **Garnet-sillimanite granite**—Light-grayish-pink or grayish-yellow, medium- to coarse-grained, xenomorphic-granular syenogranite with accessory magmatic garnet, sillimanite, and minor biotite (less than 5 percent total).

Mapped in small area west of Copeland Mountain, but also occurs in smaller (unmapped) masses northeastward to Bluebird Lake, and at Mt Lady Washington (Cole, 1977; Braddock and Cole, 1990)

Yd **Mafic dikes and plugs**—Black to dark- or light-gray, fine- to medium-grained diabasic rocks of basaltic or andesitic composition that form a north- to northwest-trending dike swarm throughout much of northern Front Range (Hepp, 1966). Plagioclase forms phenocrysts in most dikes and matrix consists of finer-grained feldspar, hornblende, pyroxene, and magnetite, or secondary minerals related to metamorphism or hydration alteration. The dike swarm cuts 1,430 Ma-old Sherman Granite (Zielinski and others, 1981) and is cut and deformed by 1,420 Ma-old Silver Plume-type granite (Cole, 1977). Unit also includes small plugs of gabbro, quartz diorite, and granodiorite in Glen Haven area (Bucknam and Braddock, 1989). Dismembered, lensoid pieces of these dikes are contained in rotated biotite gneisses in upper, flat-lying parts of Longs Peak-St Vrain batholith, attesting to extreme plastic deformation that accompanied emplacement of viscous granite magma (Cole, 1977)

YgH **Granite of Hagues Peak**—Tan, coarse-grained to very coarse grained porphyritic biotite granite. Contains conspicuous flow-aligned phenocrysts of microcline 2–8 cm long; 10–15 percent biotite. Granite of Hagues Peak is crosscut by granite of Longs Peak (YgLP) and yields a whole-rock Rb-Sr isochron age of 1,480 Ma (C.E. Hedge, written commun., 1977, cited in Braddock and Cole, 1990)

Yqd **Quartz diorite**—Fine- to medium-grained, medium- to dark-gray, locally gneissic rock with accessory hornblende and (or) biotite (20–30 percent total). Scattered porphyroblasts of microcline are common. Rock forms small masses within granite of Hagues Peak (YgH), along Highway 34 west of Estes Park, near Mt Cairns, and at Pawnee Lake; probably intruded near same time as granite of Hagues Peak

Mesoproterozoic and (or) Paleoproterozoic Intrusive Rocks

YXp **Pegmatite**—White, medium-grained to very coarse grained pegmatite with accessory biotite, muscovite, garnet, and (or) tourmaline. Most pegmatites probably intruded along with granite of Longs Peak (YgLP) but some are deformed and spatially related to older Boulder Creek Granodiorite (XgdB); reliable field criteria for distinguishing pegmatite ages have not been established

Paleoproterozoic Intrusive Rocks

XgdB **Boulder Creek Granodiorite**—Light- to medium-gray, medium- to coarse-grained, weakly to strongly foliated granodiorite and monzogranite containing 5–15 percent biotite and 0–5 percent hornblende; phenocrysts of microcline (1–4 cm long) common (Gable, 1980b). Age of Boulder Creek Granodiorite in the Boulder Creek batholith was determined to be 1,664 ± 40 Ma (Rb-Sr isochron; Peterman and others, 1968); U-Pb zircon analyses suggest true age of intrusion was 1714 ± 5 Ma (Premo and Fanning, 2000) and that some monzogranitic phases may be significantly younger (ca. 1410 Ma; Premo, written commun., 2004). Similar rock was emplaced north of Lake Granby, and east of Granby in the Strawberry Creek area, but is separately identified as unit Xgd

Xgd **Granodiorite**—Light- to medium-gray, medium- to coarse-grained, weakly to strongly foliated granodiorite and monzogranite containing biotite and variable amounts of hornblende (10–15 percent total). Rocks are physically and mineralogically similar to Boulder Creek Granodiorite but were emplaced in bodies separate from the Boulder Creek batholith

Xg **Biotite monzogranite**—Light- to medium-gray, fine- to medium-grained, massive to gneissic rock that weathers buff to orangish buff. Contains accessory biotite (5–10 percent) and secondary muscovite. Forms small bodies within Boulder Creek batholith

XjT **Trondhjemite of Thompson Canyon**—Light-gray, very fine grained and porphyritic or medium-grained and equigranular, biotite-bearing (less than five percent) leucocratic trondhjemite; forms thick tabular intrusions that are largely concordant to relict bedding in enclosing metasedimentary rocks; some locales display weak igneous biotite foliation parallel to trends of younger cross folds in the region (Braddock and Cole, 1979). Age of emplacement is 1,726 ± 15 Ma (Barovich, 1986)

Paleoproterozoic Metamorphic Rocks

Biotite-bearing schist and gneiss units (Xb, Xbq, Xbk, Xbp) are metasedimentary rocks; hornblende gneiss units (Xh) and granitic-gneiss units (Xf, Xfb, Xfcq) were probably volcanic rocks prior to metamorphism. The biotite-bearing schists vary in metamorphic grade and minerology as shown by the metamorphic-mineral zone boundaries. Pelitic rocks south of the Moose Mountain shear zone are migmatitic and are metamorphosed to a higher grade than the M metamorphic zone boundary. The age of metamorphism is 1,713 ± 30 Ma (Rb-Sr isochron; Peterman and others, 1968) and 1,715–1,690 Ma (U-Pb zircon; Premo and others, 2007), similar to the emplacement age for the Boulder Creek Granodiorite (Premo and Fanning, 2000). The primary age of interbedded volcanic rocks ranges between about 1,790 Ma–1,760 Ma (U-Pb zircon; Premo and others, 2007), consistent with model Sm-Nd ages of about 1,800 Ma for formation of the continental crust in northern Colorado (DePaolo, 1981)

Xb **Biotite schist and gneiss**—Conspicuously banded rock marked by alternating layers of contrasting composition that probably reflect original sedimentary layering, as well as effects of metamorphic segregation and partial melting. Unit is typical of highest grade metamorphic terranes, where coarse sillimanite is common and partial-melt textures are widespread. Dark-gray to black, medium- to coarse-grained layers are rich in biotite, sillimanite, and magnetite, and locally contain high-grade cordierite and (or) garnet (Gable and Sims, 1969; Cole, 1977, 2004b). Irregular quartzofeldspathic layers, bordered by selvages rich in biotite, sillimanite, and oxide minerals, formed in place during partial melting at peak of metamorphism. This unit is equivalent to units Xbq and Xbk, combined, which could not be mapped separately in higher metamorphic-grade terranes (Cole, 1977; Braddock and Cole, 1990)

Xbq **Quartzofeldspathic mica schist**—Mica-poor schist interbedded with quartzofeldspathic metasandstone. Contains interbedded knotted mica schist and sparse, thin beds of granule-sized metaconglomerate. Locally displays crossbedding. Porphyroblastic minerals preferentially located in mica-rich layers

Xbk **Knotted mica schist**—Mica-rich biotite-muscovite schist or muscovite-chlorite phyllite that displays 0.1–0.5 in porphyroblasts or clots of metamorphic porphyroblast minerals that exhibit a lumpy or knotted outcrop appearance. Contains interbedded quartzofeldspathic mica schist and sparse, thin beds of granule-sized metaconglomerate

Xbp **Porphyroblastic biotite schist**—Muscovite- or sillimanite-rich knotted mica schist that is characterized by large biotite porphyroblasts (typically 0.3 in diameter)

Xbf **Biotite gneiss and granitic gneiss, interlayered**

Xbh **Biotite gneiss and hornblende gneiss, interlayered**

Xh **Hornblende gneiss and amphibolite**—Dark-gray, greenish-gray, fine- to medium-grained, weakly to strongly layered hornblende-plagioclase gneiss and hornblende-biotite schist, locally interlayered with massive amphibolite. Contains thin layers and pods of white to light-green calc-silicate gneiss

Xf **Granitic gneiss**—Light- to medium-gray, fine- to coarse-grained, weakly to strongly foliated rock composed primarily of quartz, oligoclase, and biotite. Microcline is generally present but amount varies from sparse to abundant; hornblende, garnet, and rare sillimanite locally present. Rocks have been described as gneissic granite, gneissic alaskite, gneissic quartz monzonite, or gneissic granodiorite. Chiefly occurs in a large irregular mass south and southeast of Grand Lake, and as folded conformable sheets within metasedimentary rocks near Thunderbolt Peak, west of the Indian Peaks Wilderness Area

Xfb **Biotite granofels**—Light- to medium-gray, fine- to medium-grained, granoblastic to weakly schistose rock, commonly contains garnet and sillimanite; locally interlayered with minor biotite gneiss (Xb) and amphibolite (Xh). Unit only occurs in southern Never Summer Mountains (O'Neill, 1981)

Xfcq **Granitic gneiss, calc-silicate gneiss, and quartzite, interlayered**—Highly variable assemblage of interlayered gneisses composed of layers, an inch to several feet thick, of some or all of the following rocks: quartzite, marble, calc-silicate gneiss, microcline-rich gneiss, amphibolite, granitic gneiss, and amphibole-microcline-quartz gneiss. These interlayered gneisses are contained within more widespread biotite gneisses. Unit only occurs southwest of Meadow Creek Reservoir along southern border of quadrangle and near Glen Haven

Geologic Setting

The area covered by the Estes Park 30′ × 60′ quadrangle displays a diverse assemblage of rocks and surficial units that range in age from Paleoproterozoic to Quaternary (fig. 3). The geology records several major rock-forming events and deformations that are typical of the Southern Rocky Mountains of Colorado.

The core of the Front Range consists of igneous and metamorphic rocks that formed in the Paleoproterozoic between about 1,770 and 1,680 Ma. They were subsequently intruded and locally intensely deformed again during emplacement of widespread biotite granite at about 1,400 Ma. The east flank of the Front Range is marked by ridges of upturned sedimentary rock in the foothills-hogback belt, where tilting reflects Late Cretaceous-Paleogene (Laramide) mountain uplift. The sediments in these strata were initially deposited as early as the Pennsylvanian Period in response to the uplift and erosion of the older Ancestral Rocky Mountains. The oldest sediments were deposited adjacent to these now-eroded mountains by braided fluvial systems, and were succeeded by eolian, deltaic, marine, lagoonal, and other fluvial deposits that accumulated slowly through Jurassic time. By Early Cretaceous time, the dominant depositional environment was a fluvial delta system and lagoons on the margin of a vast inland sea; subsidence led to marine incursion and deposition of very thick shales and local limestones and sands. All of these Phanerozoic rocks underlie the hogback belt and the Colorado Piedmont areas of the eastern part of the Estes Park quadrangle.

A major regional tectonic event known as the Laramide orogeny impacted this area beginning in latest Cretaceous and led to the initial uplift of the Front Range in its modern configuration (Tweto, 1975). The eastern margin of the range rose on west-dipping, high-angle reverse faults (many are blind at the present level of erosion) and secondary east-dipping reverse faults that border en echelon tilted blocks of the Proterozoic basement. The western margin of the Laramide Front Range is marked by more complex, east-dipping thrusts that place Proterozoic crystalline rocks above steep and overturned Phanerozoic sedimentary rocks as young as Paleocene, exposed in the westernmost part of the quadrangle (O'Neill, 1981).

The Late Cretaceous was also marked by intrusive activity and hydrothermal mineralization within the Colorado Mineral Belt, which affected the Jamestown, Gold Hill, Ward, and the Audubon-Albion stock areas in this quadrangle (Lovering and Goddard, 1950).

The Laramide orogeny, probably assisted by intrusions and thermal effects of the Colorado Mineral Belt, led to withdrawal of the inland sea and exposed the Late Cretaceous Front Range to erosion. Fluvial deposits accumulated in flanking basins on both the east and west margins of the Laramide highlands. Continued erosion and deep weathering during the hot, humid climate of the Eocene and Oligocene Epochs reduced the height of the Laramide uplifts and produced residual ridge crests and large areas of subdued rolling topography that are reflected today in broad areas with nearly concordant hilltop elevations west of the foothills-hogback belt.

Renewed intrusion and volcanism during the Oligocene Epoch is recorded by stocks, flows, and ash-flow tuffs of the Braddock Peak intrusive-volcanic complex near the northwestern corner of the quadrangle. This younger magmatic activity was accompanied by high-angle normal faulting, both of which might have been manifestations of broad regional uplift related to crustal thinning and rise of aesthenospheric mantle (Eaton, 1986, 1987) and graben faulting in the northern extents of the Rio Grande rift. Regional uplift led to erosion across the Front Range and deposition of blanketing upper Miocene and lower Pliocene gravels and sands east and north of this quadrangle (Steven and others, 1997).

The Pliocene and younger history of this area is dominated by renewed uplift and erosion (Steven and others, 1997), accompanied by significant climate change to cooler and wetter conditions (Fleming, 1994; Chapin and Kelley, 1997; Bluemle and others, 2001; Williams and Cole, 2007). The drainage system of the South Platte River to the east has gradually eroded through the blanketing Miocene gravels and excavated the hogback ridges flanking the Front Range. Streams in the lower parts of the mountain block have been deeply incised into the Proterozoic crystalline core of the range.

Climate cooling that began during the Pliocene Epoch continued to the full glacial-interglacial conditions that characterize the Pleistocene and Holocene (Bluemle and others, 2001). Two major phases of mountain-valley glaciation are well recorded by high-altitude deposits within the Estes Park quadrangle (Madole and others, 1998). The spectacular alpine scenery along the Continental Divide reflects both glacial sculpting in cirques and downvalley transport of rubble by the tongues of ice.

Paleoproterozoic Layered Rocks

Proterozoic igneous and metamorphic rocks exposed in the Front Range uplift include the oldest rocks in this part of the North American cratonic basement. The metamorphic schists and gneisses were chiefly derived from marine sediments that were deposited south of the older (Archean) craton that is recognized just north of the Colorado-Wyoming State line (Reed and others, 1987; Aleinikoff and others, 1993; Reed, 1993). These metasediments, and interlayered metavolcanic rocks (erupted at about 1,760 Ma to 1,780 Ma), were progressively metamorphosed at about 1,715 Ma to 1,700 Ma (Premo and others, 2007) during the time they were being folded and intruded by calc-alkaline granodiorite and granite (Routt Plutonic Suite of Tweto, 1987).

The Estes Park quadrangle preserves an unusually complete record of this progressive regional metamorphism (from chlorite-sericite phyllite to sillimanite-biotite schist to partially melted biotite gneisses with residual high-grade cordierite and (or) garnet). Metasedimentary rocks in the Front Range in general contain sillimanite and show evidence

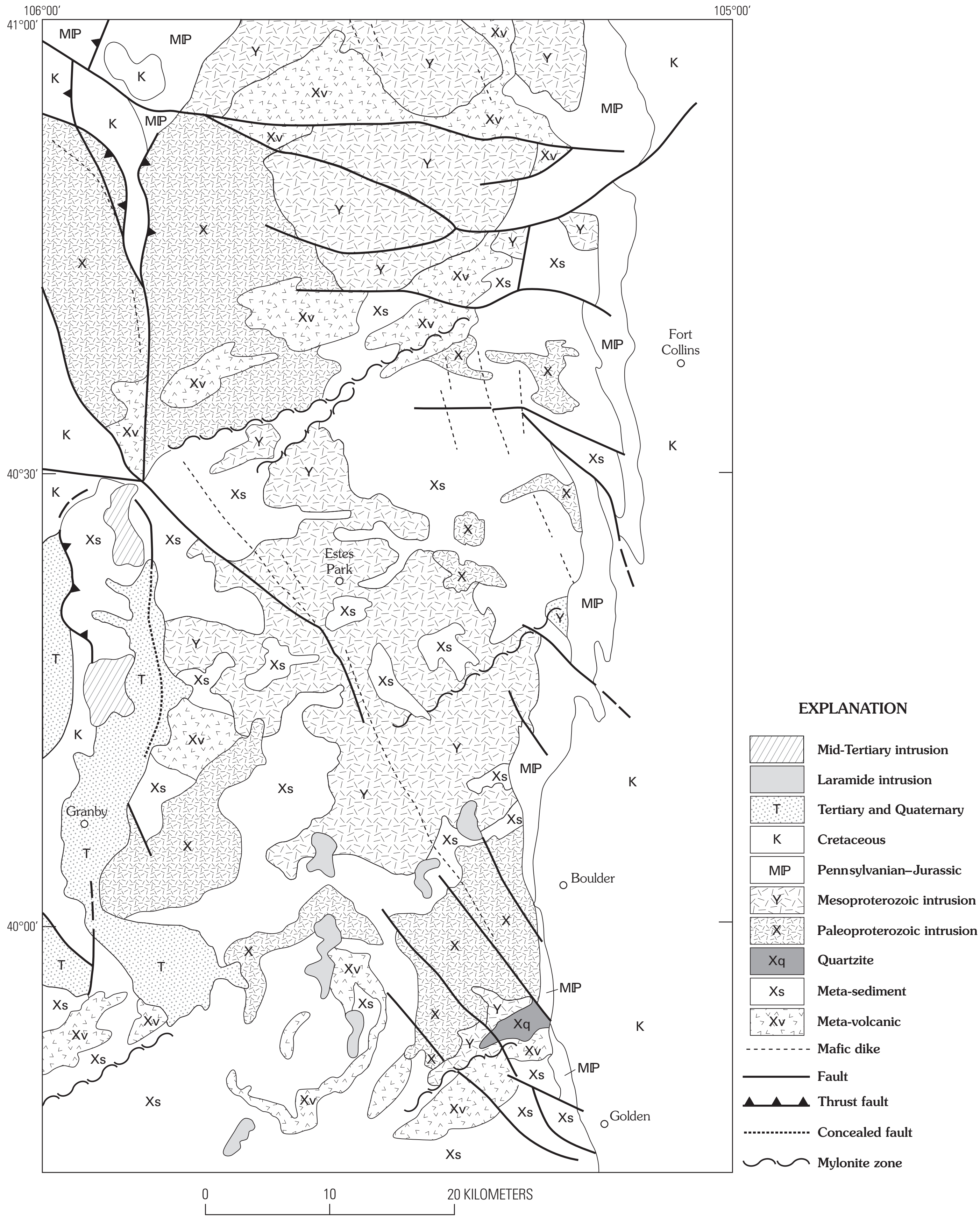


Figure 3. Map showing regional geologic setting of the northern Front Range, Colorado.

of variable degrees of partial melting (Reed and others, 1987). In the area of the Big Thompson River canyon east of Drake, sillimanite-free assemblages are preserved and the grade of metamorphism decreases eastward to greenschist-facies phyllites (Braddock and Cole, 1979).

The most common metamorphic unit in the quadrangle is a metasedimentary rock consisting chiefly of quartz, feldspar, and micas. Original sedimentary structures preserved in low-grade rocks near the mouth of the Big Thompson River canyon (Braddock, Calvert, and others, 1970) include graded bedding, crossbedding, scour-and-fill structures, thin conglomerate lenses, and current-bedding lineations (Braddock, 1970). These structures, and the generally fine-grained nature of the parent sediments, are consistent with submarine-fan depositional environments. The two most common rock types are distinguished from each other by the proportions of original clay/sand: the mica-rich variety contains abundant mica and was probably derived from shale (Xbk); the quartz-feldspar-rich variety contains less than about 15 percent mica and was probably derived from sandstone and siltstone (Xbq). Thin, persistent beds of very mica-rich rock distinguished by conspicuous porphyroblasts of biotite (Xbp) are mapped within the metasedimentary sequence north of the Big Thompson River canyon.

The distinctions among these varieties of metasedimentary rock diminish in the central and western parts of the quadrangle due to widespread partial melting during extreme metamorphism. For this reason, all metasedimentary rocks west of the longitude of Estes Park (approximately) are mapped as biotite gneiss (Xb).

Mafic schists and gneisses are locally prominent in the Paleoproterozoic metamorphic terrane of the Front Range. They generally form layers that are conformable with the surrounding metasedimentary rocks, are chemically similar to basalt and andesite, contain local relict phenocrysts, and are interpreted to have originated from volcanic flows. They typically occur as hornblende schist, hornblende gneiss, or amphibolite (Xh). Epidote-bearing calc-silicate schist and gneiss are rare but these distinctive rock types are locally present in this part of the Front Range. They commonly occur in association with the mafic metavolcanic rocks and may have formed from sediments eroded from them.

Quartzofeldspathic, leucocratic gneisses are conspicuous elements of the metamorphic terrane in parts of the Estes Park quadrangle, especially north of Lake Granby, east of Monarch Lake, and in the southern Never Summer Mountains. These gneisses typically display compositional banding marked by differing proportions of quartz, feldspar, and biotite or by variations in grainsize, and their bulk compositions are similar to granite, granodiorite, or tonalite. These felsic gneisses are commonly interlayered with mafic metavolcanic gneisses (O’Neill, 1981; Schroeder, 1995b) or form conformable layers within metasedimentary units (Pearson, 1980), and are interpreted to have formed from felsic volcanic rocks. Similar granitic gneisses in the area north of the Cache la Poudre River (also interlayered with hornblende gneiss) contain magmatic zircons that are among the oldest in northern Colorado, consistent with primary volcanic origin of these granitic gneisses at about 1,770 ± 10 Ma (Premo and others, 2007).

Paleoproterozoic Intrusive Rocks

Proterozoic intrusive rocks are common in the Estes Park quadrangle and consist of calc-alkaline granitoids ranging from monzogranite to tonalite to trondhjemite in composition. These rocks belong to the regional Routt Plutonic Suite of Tweto (1987). Their contacts with enclosing metamorphic rocks are broadly conformable to compositional layering and commonly lobate, suggesting expansive deformation of the wall rocks during emplacement. Some contacts are locally sharply discordant, but these intrusive rocks also form crescent-shaped masses (phacoids) in the cores of wall-rock folds that suggest emplacement during regional deformation (Braddock and Cole, 1979).

The most widespread unit is the Boulder Creek Granodiorite (XgdB), which was named from its type area in the Boulder Creek batholith that extends into the south-central part of this quadrangle (Gable, 1980b). The Boulder Creek is typically a foliated granodiorite containing biotite and (or) hornblende that shows concentrations of mafic minerals in schlieren bands, aplitic-pegmatitic zones, and local feldspar porphyry textures. Gable (1980b) was able to identify irregular masses within the Boulder Creek batholith consisting of tonalitic, granodioritic, and monzogranitic compositions, but all such internal distinctions are gradational. Internal dikes are not common, but generally consist of leucocratic granitic and aplitic rocks or pegmatite. The Boulder Creek batholith was emplaced at 1,713 ± 4 Ma, based on the statistical weighted-mean of values determined from individual zircon grains (Premo and Fanning, 2000).

Rock similar to the type Boulder Creek Granodiorite (XgdB) has been mapped in three other principal locations in the quadrangle, designated as Paleoproterozoic granodiorite (Xgd). Pearson (1980) mapped a substantial pluton of foliated granodiorite east and west of Arapahoe Pass along the southern border of the quadrangle. This body is largely conformable to compositional banding in the enclosing biotite gneisses and contains large concordant sheets of the same. Schroeder (1995b) mapped a large, equant intrusive body in the Strawberry Lake quadrangle that shows generally conformable contacts with biotite gneiss around its margins and steep internal foliation; phases of this pluton are conspicuously porphyritic with microcline phenocrysts as long as 16 cm. Complex amoeboid bodies of Boulder Creek-type granodiorite were mapped in the Glen Haven quadrangle in the vicinity of Crosier Mountain by Bucknam and Braddock (1989); they are intimately intruded by, and surrounded by extensive bodies of pegmatite-aplite that were probably emplaced at Boulder Creek time.

Biotite monzogranite and syenogranite (Xg) have been mapped in several small areas of the quadrangle that differ from the Boulder Creek in composition but probably belong to the Paleoproterozoic Routt Plutonic Suite (Tweto, 1987). Many of these are small masses within the Boulder Creek batholith, referred to as Twin Spruce monzogranite ("quartz monzonite" of Gable, 1980b). Detailed zircon geochronology by Premo and Fanning (2000), however, has demonstrated that some monzogranite within the Boulder Creek batholith was emplaced much later during the Berthoud Plutonic Suite at about 1,400 Ma.

Distinctive fine-grained biotite trondhjemite (XjT) intrudes the metasedimentary section in the vicinity of the Big Thompson River canyon. This rock typically forms continuous, concordant sheets that can be traced for miles along the strike of sedimentary bedding (Braddock, Nutalaya, and others, 1970). Several stubby sills coalesce at Palisade Mountain, east of Drake, to form a cylindrical plug; surrounding sedimentary bedding is deformed outward by expansive emplacement of the trondhjemite. Braddock and Cole (1979) inferred that the trondhjemite magmas were emplaced during the waning stages of regional folding because they crosscut an older set of regional folds but contain aligned biotite in the direction of axial surfaces of the youngest set of regional folds. These observations were confirmed in more detailed work by Barovich (1986), who also determined the trondhjemite intruded at 1,726 ± 15 Ma by U-Pb zircon methods.

Paleoproterozoic Metamorphism

Metamorphic rocks of the Estes Park quadrangle record evidence of two regional metamorphic events. The first is only preserved in the lowest grade phyllites and schists and is indicated by aligned sericite, chlorite, and muscovite that probably grew from detrital clay minerals. This first metamorphism seems to have accompanied and followed the first regional deformation during which major tight folds formed within the thick pile of saturated marine sediments (Braddock, 1970; Braddock and Cole, 1979). Original clay minerals were aligned parallel to axial-plane cleavage of first-generation folds, a process that involved laminar intrusion of clay-rich seams through more granular sand-silt beds during soft-sediment deformation (Braddock, 1970). Subsequent recrystallization of these clay minerals during the earliest metamorphism produced sericite, chlorite, and muscovite whose fabric alignment mimicked the alignment of clay minerals.

The second episode of metamorphism produced the pervasive prograde mineral assemblages within the basement rocks of the Front Range and largely coincided with intrusion of the Boulder Creek Granodiorite (XgdB) at about 1,715 Ma (Peterman and others, 1968; Braddock and Cole, 1979; Hutchinson and Braddock, 1987). Metamorphic grade ranges from greenschist near the mouth of the Big Thompson River canyon (fig. 4*A*) to upper-amphibolite/partial-melting west of Drake. This higher metamorphic grade is typical of most areas in the Front Range. The lower-grade assemblages are preserved in an east-west swath that marks a fossil thermal "trough" outlined by nested isograds defined by the appearance/disappearance of zonal minerals (and mineral assemblages) from east to northwest, west, and southwest (fig. 4*A*; Braddock and Cole, 1979; Nesse, 1984). From low- to high-grade, pelitic rocks display porphyroblasts of chlorite, chloritoid, biotite, garnet, staurolite, cordierite, andalusite, and potassium feldspar that formed through sequential reactions (figs. 4*B*–*E*). Sillimanite first appears in pelites at about the position that low-grade garnet and cordierite disappear, and staurolite and andalusite disappear a short distance up-grade from the sillimanite-in line. All prograde muscovite was consumed at temperatures below the onset of melting by the reaction muscovite + quartz = potassium feldspar + sillimanite + water (Nesse, 1984).

Partial melting (summarized from Cole, 1977, 2004b) occurred in the pelitic and quartzofeldspathic metasediments that consisted of varying proportions of quartz, plagioclase, potassium feldspar, biotite, sillimanite, and magnetite. The low-melting fraction of quartz and feldspars melted first (in the presence of water) and segregated into seams and lenses (leucosomes), generally parallel to compositional banding in the host rock. These leucosomes are typically bordered by dark-colored selvages (melanosomes) that contain enriched concentrations of the non-melted fraction of the rock (chiefly biotite, sillimanite, and magnetite). This distinctive texture of leucosome-melanosome segregations within the non-melted host rock is the characteristic of migmatite, and its first appearance has been systematically mapped in the Estes Park quadrangle, forming the outermost reaction-isograd of the nested zonal isograds along the course of Big Thompson Canyon (Braddock and Cole, 1979). The fact that the migmatite-in isograd lies generally parallel to the other mineral-reaction isograds, and in the direction of inferred higher temperatures and pressures, is strong evidence that the leucosomes formed in place by partial melting and were not injected from some external magmatic source (Nesse, 1984).

Progressive partial melting is indicated by the local appearance of new porphyroblastic cordierite and (or) garnet in more strongly melted rocks (fig. 4*D*). These minerals originated through incongruent melting of biotite, a process that added iron and aluminum to the partial melt and produced magnesium-enriched cordierite or garnet (or both) in the residue, depending on bulk composition of the rock (Cole, 1977). This high-grade cordierite, and garnet to a lesser degree, replaces sillimanite and biotite (Gable and Sims, 1969; Cole, 1977, 2004b).

Regional metamorphism took place during the protracted orogenic event marked by the intrusion of Boulder Creek Granodiorite (XgdB) and by the formation of kilometer-scale folds of two contrasting trends (Braddock and Cole, 1979; Hutchinson and Braddock, 1987). The thermal maximum is interpreted to have coincided with the peak of deformation in the highest-grade rocks. The conclusion is supported by 1) partial-melt leucosome masses that are phacolithic in fold hinges and elongate parallel to fold axes (fig. 4*E*), 2) coarse sillimanite and biotite aligned with fold axes (fig. 4*C*), and 3) some potassium feldspar porphyroblasts containing helicitic inclusions of quartz and magnetite that suggest rotation during growth. This conclusion has not yet (2007) been confirmed by high-resolution geochronology.

The evidence from the lower-grade terranes is quite different. Mineral textures here indicate the thermal peak of metamorphic recrystallization traversed the rocks following the conclusion of folding. All porphyroblastic minerals, and especially andalusite, cordierite, and staurolite, are irregular and amoeboid and overprint the traces of small folds defined by shingled muscovite, biotite, quartz, and feldspars (Cole, 1977, 2004b; Braddock and Cole, 1979).

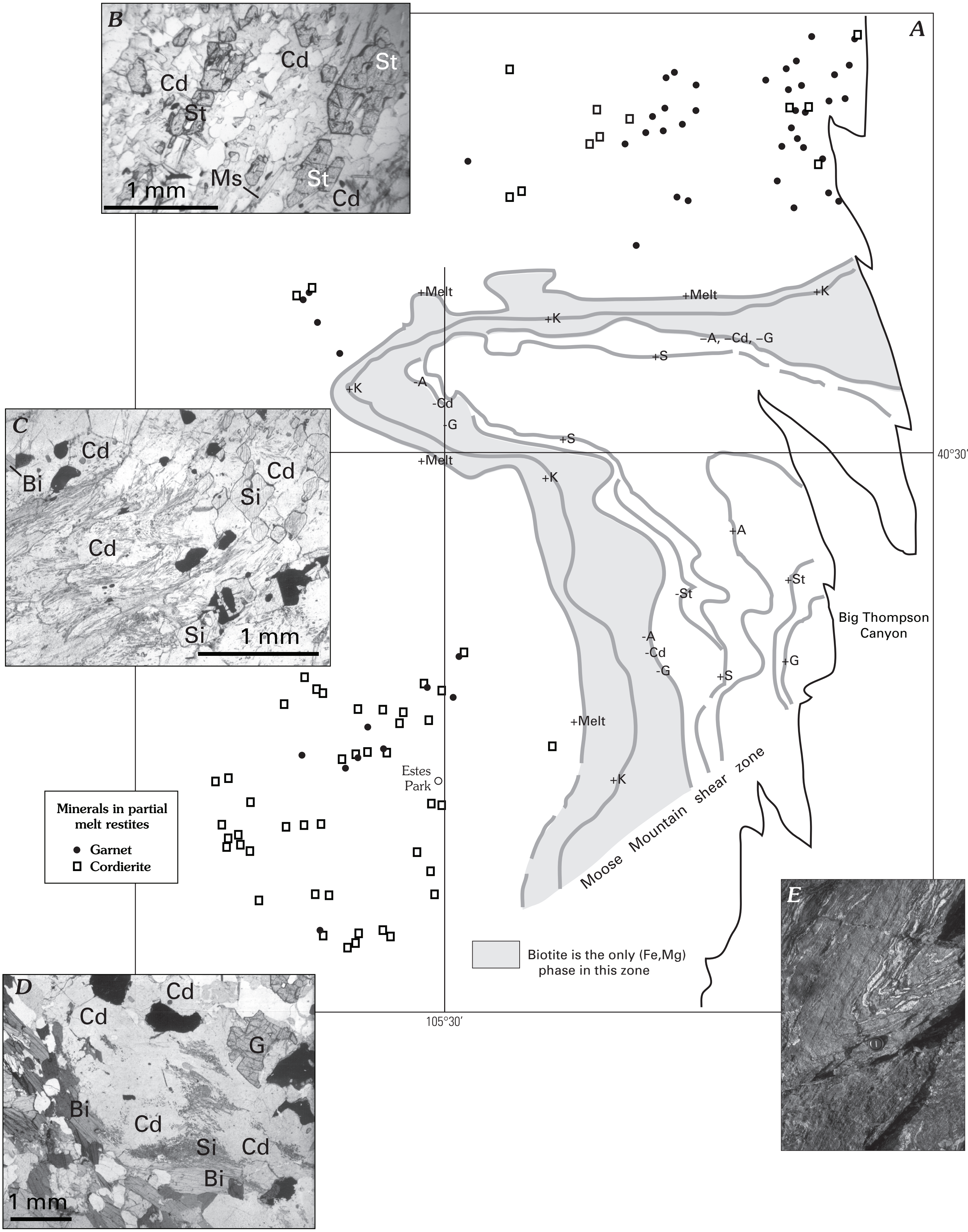
Interpreted conditions at the peak of metamorphism have been estimated by several authors. Cole (1977, 2004b) concluded that peak conditions were about 5.5 kb, 675°–725° C, and undersaturated with vapor on the basis of regional relations and experimentally calibrated phase equilibria. Nesse (1984) used mineral compositions and phase equilibria to conclude that the muscovite + quartz = potassium feldspar + sillimanite + water reaction occurred at 3–4 kb and about 650° C under water-saturated conditions (higher pressures would have prevailed under water-undersaturated conditions; Cole, 1977). Both authors conclude that pressures greater than about 6 kb could not have been attained because no kyanite has ever been reported in this region.

Detailed mineralogical and chemical studies in a few scattered localities in the Front Range have led to conclusions widely different from the above. Munn and Tracy (1992) and Munn and others (1993) reported thermobarometric calculations based on garnet, biotite, and hornblende chemistry that suggested pressures near 7 kb north of the Estes Park quadrangle. Selverstone and others (1995, 1997) assert pressures of 8 kb to 10 kb based on unpublished chemistry of garnet and plagioclase in staurolite-bearing schists near the Big Thompson River, but do not explain how such extreme conditions could have been attained in the absence of partial melting or without formation of kyanite. We discount these conclusions and suggest that either the chemical analyses were obtained from non-equilibrium assemblages (Cole, 2004b) or that the underlying chemistry-temperature/pressure calibrations are not applicable to these rocks.

Retrograde metamorphism is manifest by three distinct kinds of mineral transformations that depend on the extent of prograde metamorphism. The overall patterns of retrograde assemblages indicate that they were not influenced by intrusions of much younger plutons at about 1,400 Ma (Cole, 2004b). In the lower-grade schists, aluminous porphyroblasts are replaced by very fine grained sericite and chlorite, accompanied by tourmaline. Retrograde muscovite is common in medium- and high-grade metasedimentary rocks where it forms irregular, non-oriented porphyroblasts that replace oriented sillimanite and biotite; tourmaline and topaz also occur in this assemblage (Cole, 1977). This porphyroblastic retrograde muscovite is mostly confined to rocks near and below the migmatite-in isograd and seems to indicate it formed with water that was released from crystallizing leucosome melts that reacted with potassium feldspar and sillimanite (Cole, 1977). The highest-grade metasediments show only limited retrograde reactions, probably because most of the water released from crystallizing partial melts had already migrated to higher, cooler areas of the orogen. Small amounts of wormy andalusite are present in some migmatites, particularly those with abundant high-grade cordierite indicating high degrees of partial melting (Gable and Sims, 1969; Cole, 1977). The origin of this late andalusite is not clear, but textural evidence suggests it is related to breakdown of biotite to andalusite + magnetite under very dry conditions (Cole, 1977).

Selverstone and others (1995, 1997) and Shaw and others (1999b) have asserted that the metamorphic assemblages in this area reflect the combined effects of regional metamorphism at about 1,710 Ma and again at about 1,400 Ma. These statements appear to be based on ambiguous textural relations involving staurolite, and on numerous 1,400 Ma ages determined by Ar–Ar analysis from prograde minerals and similar U-Pb results for monazite (see also Shaw and others, 1999a). It has been known for decades that most single-mineral isotopic systems in the Colorado basement were reset due to generalized heating that accompanied intrusion of widespread granitic rocks of the Berthoud

Figure 4 (facing page). Photographs and map showing features of Paleoproterozoic metamorphism in the northern Front Range (*C*, *D*, *E* photographs, J.C. Cole, 1972–1974). *A*, Sketch map showing distribution of metamorphic-mineral isograds in pelitic rocks, marked by the first appearance (+) or disappearance (-) of andalusite (A), biotite (Bi), cordierite (Cd), garnet (G), potassium feldspar (K), sillimanite (S), or staurolite (St). Line marking appearance of partial-melt textures (+melt) indicates the western and northern extent of unmelted (low-grade) pelitic rocks. Garnet and cordierite in restite phase of partially melted gneisses (dot and square symbols) originated through incongruent melting of biotite and are compositionally and texturally distinct from garnet and cordierite in the lower-grade terrane to the east and southeast. *B*, Photomicrograph showing matrix cordierite (Cd), quartz (colorless), and muscovite (Ms) with porphyroblastic staurolite (St) in low-grade schist (W.A. Braddock photograph, 1970). *C*, Photomicrograph showing coarse, oriented prisms of sillimanite (Si) intergrown with polygonal cordierite (Cd) that replaces fine sillimanite needles and remnant biotite (Bi) in partially melted biotite gneiss. *D*, Photomicrograph showing polygonal cordierite (Cd) and irregular garnet (G) replacing biotite (Bi) and sillimanite (Si) in partially melted biotite gneiss. *E*, Outcrop photograph showing partial-melt textures (light-colored quartz-feldspar leucosomes surrounded by dark melanosomes containing unmelted biotite, sillimanite, and magnetite) in partially melted biotite gneiss. Leucosomes form rod-shaped bodies elongate parallel to fold axis due to growth during deformation. Lens cap about 2 in diameter.


A
B
C
D
E
Cd
St
Ms
1 mm
Bi
Si
G
+Melt
+K
+S
−A, −Cd, −G
-A
-Cd
-G
+A
-St
+St
+G
40°30′
105°30′
Big Thompson Canyon
Estes Park
Moose Mountain shear zone
Minerals in partial melt restites
Garnet
Cordierite
Biotite is the only (Fe,Mg) phase in this zone

Plutonic Suite (for example, Peterman and others, 1968). However, the textural and regional field evidence described above clearly shows that the sequential mineral assemblages preserved in the metasedimentary rocks formed during a single major Paleoproterozoic prograde event. The age recorded by closure of isotopic systems within individual minerals does not date the time of growth of the particular minerals.

Mesoproterozoic Intrusive Rocks

Younger Proterozoic intrusive rocks were emplaced between about 1,480–1,380 Ma (Berthoud Plutonic Suite of Tweto, 1987) and are widespread in the Estes Park quadrangle. These include a regional swarm of northwest-trending mafic porphyry dikes, batholithic masses of distinctive Silver Plume-type granite, smaller bodies of granodiorite and quartz diorite, and pegmatite. In addition, an unusually long, narrow ferrogabbro dike swarm known as the Iron Dike (Ygb) was intruded across the entire quadrangle and beyond at about 1,315 Ma (Braddock and Peterman, 1989).

The oldest magmas of the Berthoud Plutonic Suite are relatively sparse in this area and are more mafic than the abundant, younger magmas. The granite of Hagues Peak (YgH; north-central border of the quadrangle) is a biotite monzogranite to granodiorite that is distinguished by large microcline phenocrysts and as much as 15 percent biotite, both of which are flow-aligned; it was intruded at about 1,480 Ma (Braddock and Cole, 1990). Small masses of undated biotite-hornblende quartz diorite are present in the area west of Estes Park and are cross-cut by the granite of Hagues Peak (YgH) and by widespread Silver Plume-type granite.

Mafic porphyry dikes form a persistent and widespread swarm that trends north-northwest through the metamorphic rocks of the Front Range. These distinctive dikes are typically 1–5 m thick and 100 m to several kilometers long and are especially common in the areas west of Loveland and north of this quadrangle. They consist of basalt and mafic andesite with conspicuous plagioclase phenocrysts in a fine-grained groundmass with diabasic texture (Hepp, 1966; Kellogg, 1973). These dikes crosscut older batholiths of the Berthoud Plutonic Suite (Sherman Granite) and are intruded and deformed by younger batholiths of Silver Plume-type granite (Eggler, 1968; Peterman and others, 1968; Cole, 1977). Their emplacement age is constrained by the ages of these batholith granites between about 1,415 Ma and 1,405 Ma.

Silver Plume Granite is a recognized lithodemic-unit name (Lovering and Goddard, 1950; Tweto, 1987) for widespread intrusive rocks in the Front Range that form irregular batholiths, simple ovoid plutons, and dikes. The name is based on the type area at Silver Plume, Colorado (60 km south of Estes Park), where characteristic mineralogical and textural traits are displayed (Tweto, 1987). Throughout the Front Range (but outside the type area of the Silver Plume batholith), granites that possess Silver Plume characteristics have been correlated with the type rock-unit and the name Silver Plume Granite has been used in order to express the general similarities in age, texture, and composition. However, each batholith of Silver Plume-type granitic rocks conveys its own history and it is likely that each arose at somewhat different times under somewhat individual circumstances.

In the interest of clarity, and in concert with the compilers of adjacent 30′ × 60′ quadrangles, we elected to restrict use of the term Silver Plume Granite to the Silver Plume batholith in its type area. We describe correlated rock masses elsewhere in the northern Front Range as Silver Plume-type granites. Each individual intrusive complex is designated according to the historically established name of the pluton or batholith where it intruded. In this quadrangle, granite of Longs Peak (YgLP) denotes the Silver Plume-type granitic rocks within the Longs Peak-St Vrain batholith.

The granite of Longs Peak (YgLP) is fairly homogeneous in composition and mineralogy, but shows local textural and grain size variations. The most typical rock types consist of biotite monzogranite and biotite syenogranite, some of which contain accessory sillimanite and (or) garnet. The most common variety of granite of Longs Peak (YgLP) is coarsely porphyritic and contains conspicuous tabular phenocrysts of microcline, typically flow oriented (fig. 5*A*), that are 1–3 cm long and set in a fine- to medium-grained matrix of quartz, feldspars, biotite, and accessory minerals. This trachytic feldspar foliation is a mappable feature within the granite and defines internal structure formed by flow patterns of the magma as it intruded (Cole, 1977). Less common varieties are finer grained, less porphyritic, more leucocratic, and some grade into aplite-pegmatite bodies.

Anderson and Thomas (1985) studied the chemistry, mineralogy, and isotopic characteristics of Silver Plume-type granites in this area. They noted the presence of muscovite in some samples and interpreted it to be a primary, magmatic phase based (presumably) on the fact that biotite and muscovite are “intergrown.” We conclude otherwise that the muscovite is a subsolidus mineral formed by breakdown of magmatic sillimanite and potassium feldspar (to muscovite + quartz), and by the breakdown of biotite to muscovite and magnetite (Cole, 1977; Braddock and Cole, 1990). Our conclusion is based on textural and spatial criteria. The magmatic biotite is generally flow-aligned parallel to feldspars in the rock, but subsolidus muscovite is everywhere randomly oriented. In addition, this late muscovite is preferentially located near the external margins of the major batholith and is essentially absent from the center (Cole, 1977).

Almost all of the Silver Plume-type granite in the Estes Park quadrangle was emplaced within the structurally complex Longs Peak-St Vrain batholith (Boos and Boos, 1934; Cole, 1977). This body extends from the Left Hand Creek drainage on the south to just beyond the northern quadrangle boundary, and from the Colorado River on the west to beyond the unconformity with the Phanerozoic sediments on the east. The body is irregular in three-dimensional form but spans approximately

30 mi by 35 mi and comprises nearly half of the Proterozoic rocks exposed in the quadrangle.

External contacts of the Longs Peak-St Vrain batholith are discordant to wall rock foliation in detail, but broadly conformable. Outer contacts tend to be relatively steep and lobate. Trachytic foliation within the granite defines local antiformal patterns that are separated by synformal screens of metamorphic rock (fig. 5*C*, 10 following). Cole (1977) attributed these structural patterns to gravity-induced sinking of denser metasedimentary rocks between buoyant, rising lobes of the viscous magma near the time of pluton consolidation.

Contacts with wall rock in the upper most parts of the Longs Peak-St Vrain batholith dip very gently and are strongly concordant with wall rock foliation (figs. 5*C*–*E*). The intrusion geometry is well displayed in the glaciated valleys on both sides of the Continental Divide, where numerous flat-lying sheets of the light-colored granite are separated by conformable tabular masses of dark-colored biotite gneiss.

Cole (1977) argued that these contact relations indicate the magma was nearly as viscous as the wall rock during intrusion, and that both of them deformed together as the buoyant magma rose into the middle crust (fig. 5*B*). The flat-lying foliation of the metamorphic wall rocks is the result of bodily rotation (from pre-existing near-vertical attitudes) during emplacement of the granite. Remnants of the mafic porphyry dikes that were regionally emplaced as sub-vertical dike swarms just prior to the Longs Peak-St Vrain batholith are preserved only as dismembered, flattened tabular slabs within the enclosing biotite gneisses. Reconnaissance paleomagnetic investigation of these dismembered dike segments showed that the formerly vertical dikes had experienced large angular rotations in the deformed contact zone of the batholith (Cole, 1977).

The Silver Plume-type granites are representative of a suite of similar granitic intrusive rocks that were emplaced in the North American craton at about 1,400 ± 50 Ma along a broad belt between southern California and Labrador. These granites are compositionally similar, chiefly biotite monzogranite or syenogranite, and are either metaluminous or slightly peraluminous (Anderson, 1983). Most have initial strontium-isotope ratios that are similar to the calculated 1,400-Ma mantle and thus do not seem to have assimilated large amounts of more-radiogenic wall rock. Many share the characteristic porphyry texture with the Silver Plume or the rapakivi texture of mantled feldspar phenocrysts. They generally have simple crosscutting relations with wall-rock structures (in regional context), and Anderson (1983) referred to them as “anorogenic” granites due to the general absence of evidence for coeval regional deformation. These granites do not seem to have formed through melting above a descending oceanic-crustal slab because they lack the compositional diversity typical of that environment, few basalts are associated with them, and no supportive plate-tectonic context has been identified for this part of North America at 1,400 Ma (Anderson and Cullers, 1999). We believe that their widespread occurrence, compositional homogeneity, and primitive isotopic character indicate they formed as a result of fundamental changes in mantle structure (possibly related to mineralogical phase changes) at 1,400 Ma (Cole, 1977; Anderson and Bender, 1989).

Selverstone and others (1997, 2000) have argued instead that the regional stress field during Silver Plume Granite emplacement was compressional with a shortening direction oriented north-northwest. We concur with Cole (1977) and Anderson and Cullers (1999) that regional stress was neutral or slightly extensional during intrusion. The evidence from this area (and the contrasting interpretations) are presented below in the section on Mesoproterozoic structure.

The youngest Proterozoic rock in the Estes Park quadrangle is the Iron Dike (Ygb), a distinctive intrusion swarm that consists of iron-rich augite ferrogabbro. Its chemistry was studied in some detail by Wahlstrom (1956) and Cole (1977) described its form, contact relations, and mineralogy. Braddock and Peterman (1989) reported an emplacement age of 1,316 ± 50 Ma based on a mineral-whole-rock Rb-Sr isochron. The Iron Dike consists of a swarm of north-northwest trending sub-vertical dikes that can be traced from just west of Boulder through the Gold Hill area to Allens Park, and then through Rocky Mountain National Park and Trail Ridge to the northern quadrangle boundary at Long Draw Reservoir. Further mapping to the north shows that this swarm continues along the east side of the Medicine Bow Mountains as far as the Wyoming-Colorado State line (Pearson and others, 1982; Braddock and Cole, 1990). The combined length of the swarm is approximately 95 mi. Thickness ranges from a few feet or less in some of the thinner dikes to as much as 130 ft. Contacts with the Proterozoic wall rock are typically chilled within a few inches of the margin (Cole, 1977). The Iron Dike (Ygb) was hot enough during intrusion to cause local grain-contact melting in the Silver Plume-type granite (Braddock and Peterman, 1989).

Kimberlite

Three small bodies of kimberlite intrusive rock (Dk) are present in the Estes Park quadrangle. Other kimberlite pipes and dikes have been mapped and explored in the northern Front Range between the Colorado-Wyoming State line and Boulder (Chronic and others, 1969; Smith, 1979; Eggler and Braddock, 1988; Kellogg, 1973). They typically contain abundant phlogopitic biotite, pyrope garnet, chromian diopside, and widespread carbonate and sheet-silicate alteration products. Some pipes in the State-line area have yielded small quantities of industrial-grade diamond, but none has been recovered from the Estes Park quadrangle.

Fossiliferous Cambrian through Silurian limestone blocks are contained in some of the kimberlite pipes in the State-line area, and initial fission-track studies supported a Devonian age of emplacement (Naeser and McCallum, 1977). More recent work by Lester and others (2001) shows that at least

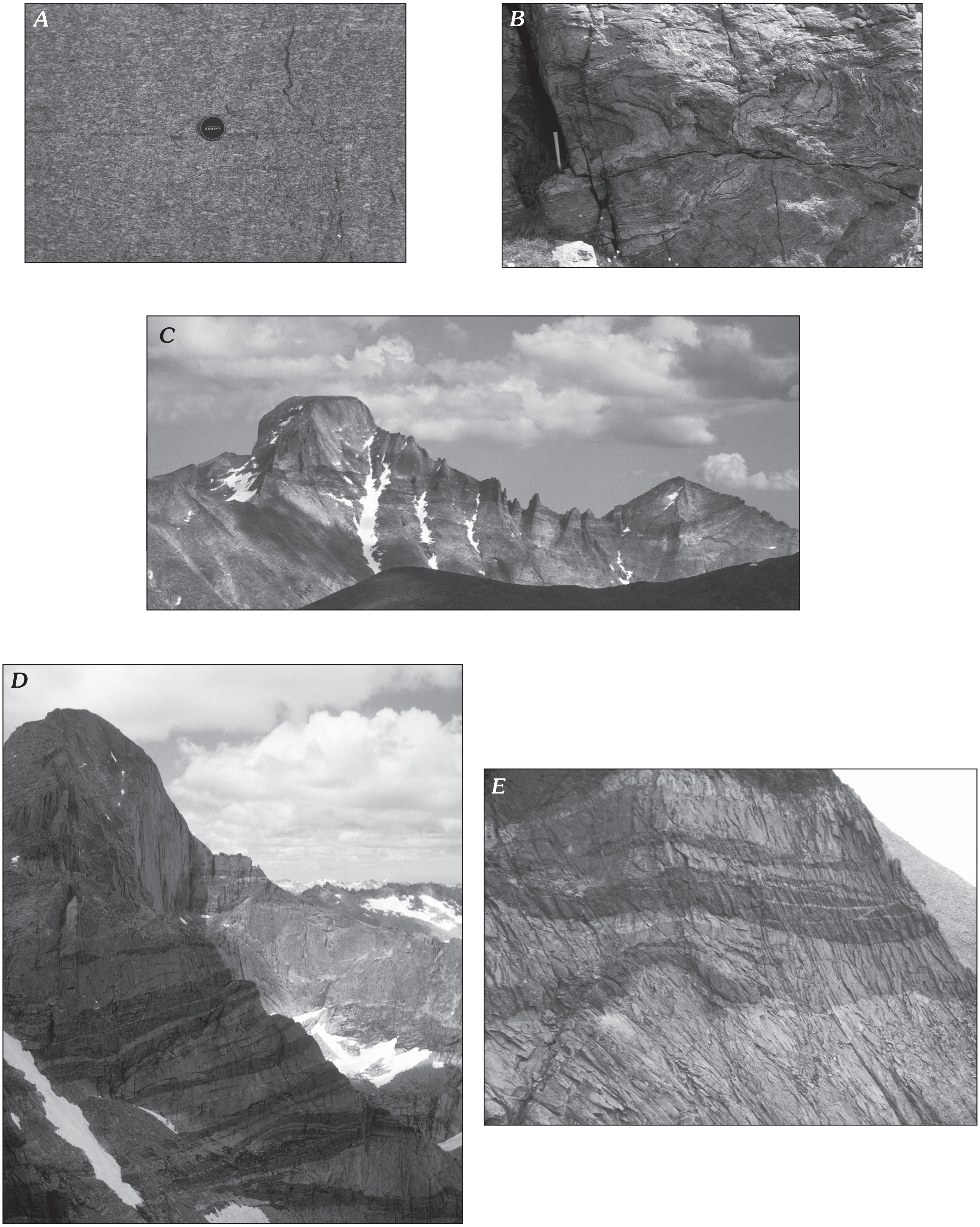

A
B
C
D
E

Figure 5 (previous page). Photographs showing internal structures and intrusive relations of the Longs Peak-St Vrain batholith (J.C. Cole photographs, 1972–1975). *A*, Outcrop photograph showing typical texture of granite of Longs Peak (YgLP) marked by strong flow-alignment of feldspar phenocrysts. Lens cap about 2 in diameter. *B*, Outcrop photograph showing sills of light-colored granite of Longs Peak (YgLP) that were folded along with the enclosing, banded biotite gneisses. Continuity of structure indicates folding took place during intrusion of granite. Hammer handle at left is about 1 ft long. *C*, Photograph of the west face of Longs Peak, Keyboard of the Winds, and Pagoda Mountain showing banded structure of intrusive granite sheets (YgLP, gray) separated by tabular sheets of biotite gneiss (Xb, dark). Arches and swales outlined by the biotite-gneiss layers show apparent folding that reflects rising, buoyant magma and sinking, dense biotite gneiss during intrusion of the Longs Peak-St Vrain batholith. *D*, Photograph looking west from Mt Meeker ridge toward the east face of Longs Peak (summit at left) showing detailed pattern of numerous sills of granite (YgLP, gray) intruded along foliation planes in biotite gneiss (Xb, dark). *E*, Photograph of steep rock wall on south ridge of Pagoda Mountain showing tabular structure formed by subhorizontal sheets of granite (YgLP, light and medium gray) that intrude each other and intrude biotite gneiss (Xb, dark). The arch in the lower-middle part of the image was formed by a finger-like extension near the margin of a tabular sheet of granite (light gray, lower part of image).

the kimberlites near Boulder and in the Cache la Poudre River drainage (north of this quadrangle) were emplaced at about 700 Ma to 600 Ma during the Neoproterozoic. Phlogopite from the kimberlite dike located about 2 mi south of Estes Park produced an Rb-Sr isochron age of 404 ± 18 Ma (Early Devonian; Smith, 1979).

Pre-Cenozoic Sedimentary Rocks

Paleozoic and Mesozoic sedimentary rocks are exposed on both the east and west flanks of the Front Range uplift in the Estes Park quadrangle. Excellent summaries of detailed investigations of the Phanerozoic units are available for this area, including Lee (1927), Scott and Cobban (1965, 1986), and Weimer (1996). These papers and many of the source geologic maps form the basis for the following short summations.

The oldest sediments deposited on the eroded and weathered Proterozoic crystalline rocks are fluvial sandstones that reflect uplift and erosion during the Ancestral Rocky Mountains orogeny in Pennsylvanian time. The Pennsylvanian and Permian Fountain Formation (PIPf) (arkosic conglomerate and sandstone; 1,100 ft thick) records the initial deposition along the eastern flank of the uplift. The Fountain Formation is overlain by Permian Ingleside Formation (Pi) (planar-bedded sandstone; 50 to 200 ft thick), Owl Canyon Formation (Po) (red siltstone; pinches out south of Little Thompson River; maximum 250 ft), and Lyons Formation (Pl) (eolian crossbedded sandstone and fluvial sand; 60 ft thick). The Lyons is overlain by Permian and Triassic Lykins Formation (ŦPl) that comprises about 550 ft of red siltstone and fine sandstone, with limestone and gypsum beds (Forelle Limestone Member and other units) near the base. The Upper Triassic Jelm Formation (in unit JŦmj) overlies the Lykins and consists of about 100 ft of crossbedded fine sandstone.

The west flank of the modern Front Range remained positive through Pennsylvanian and Early Permian time. The oldest sediments deposited there (northern Never Summer Mountains and Granby valley areas) are the Chugwater Formation (ŦPc) of Permian and Triassic age (including Forelle Limestone Member and probable upper Satanka Shale), roughly equivalent to the Lykins Formation (ŦPl) on the east flank.

By Middle Jurassic time, the Ancestral Rocky Mountains uplift had been eroded to sea level in this area. Fine-grained, crossbedded sandstone and laminated siltstone of the Sundance Formation (in unit Jms), probably equivalent to the Canyon Springs Sandstone Member, is preserved on both the east (about 15 ft thick) and west (about 125 ft thick) flanks of the Front Range. The Upper Jurassic section is represented by about 300 ft of Morrison Formation (Jm), consisting of variegated fluvial claystone, siltstone, and thin sandstones on both sides of the Front Range.

The vast, shallow Western Interior seaway flooded the center of North America beginning in Early Cretaceous time. The marginal-marine Dakota Group (Kd) of sandstone, shale, and sparse limestone comprises the Lower Cretaceous section across the whole of the Estes Park quadrangle. The disconformable base of the Dakota Group is overlain by pebbly sandstones of the Lytle Formation (about 60 ft thick) representing deposition on a channeled floodplain. The overlying South Platte Formation contains the transgressive Plainview Sandstone Member at its base (30 ft thick), deep-water shales equivalent to the Skull Creek Shale (about 90 ft thick), and beach and tidal-flat sands equivalent to the Muddy (or “J”) Sandstone at the top (40 to 100 ft thick).

Combined subsidence and sea-level rise near the end of Early Cretaceous time led to the return of full marine conditions that persisted throughout most of the Late Cretaceous. Shale, thin-bedded limestone, and sparse sands accumulated in the Western Interior seaway across this entire area and are known (from bottom to top) as the Lower Cretaceous Mowry Shale (10 ft thick) and the overlying units of the Benton Shale (Kbm), consisting of Upper Cretaceous Graneros Shale (160 ft thick), Greenhorn Limestone (260 ft thick), and Carlile Shale (80 ft thick). The overlying Niobrara Formation (Kn) consists of the Fort Hays Limestone Member (15 ft thick) and the Smoky Hill Shale Member (65 ft thick).

Increased rates of subsidence are recorded by the Upper Cretaceous Pierre Shale (Kp) across the Estes Park quadrangle. Subsurface data show that the Pierre is approximately 5,500 ft thick in North Park (northwest of this quadrangle) and 6,800 ft thick in the deepest parts of the Denver Basin east of Boulder, all of which was deposited in about 14 million years (Obradovich and Cobban, 1975; Obradovich, 1993). The isopach patterns of Late Cretaceous deposition suggest the accelerated subsidence may have been caused by crustal flexure, a possible harbinger of the Laramide Front Range uplift. From base to top, the Pierre Shale (Kp) is compiled on this map as the lower shale member (Kpl); Hygiene Sandstone Member (Kph); middle shale member (Kpm); combined Rocky Ridge Sandstone, Larimer Sandstone, and Richard Sandstone Members (Kpr); upper shale member (Kpu); and upper transition member (Kpt) (Scott and Cobban, 1965, 1986).

The Western Interior seaway began to recede in Late Cretaceous, and its withdrawal in this area is marked by the strandline marine deposits and coeval, interfingering, and younger deposits of coastal plain, coal-forming swamp, and delta-plain environments (Roberts, 2005). The shore-line sands are the Fox Hills Sandstone (Kfh), which consists of shingled sand bodies representing repeated minor transgressions and regressions along the seaward edge of a large delta complex (Davis and Weimer, 1976; Weimer, 1996); aggregate thickness is as great as 330 ft in the Erie quadrangle (Colton and Anderson, 1977; Roberts, 2005). The overlying coastal-plain deposits comprise the Laramie Formation (Kl) that contains numerous seams of commercially extracted coal.

Laramide Synorogenic Sedimentary Rocks

The landscape of much of the Rocky Mountains area changed dramatically as the Western Interior seaway receded near the end of Cretaceous time. The outlines of the modern mountain ranges in this part of Colorado were established when subduction-related buckling initiated uplift on mountain-scale crustal blocks (Dickinson and Snyder, 1978). The deformation (tilting and uplift) began at somewhat variable times in different places (Kluth and Nelson, 1988). The end of marine deposition and the transition to coastal and subaerial environments is generally taken to mark the onset of the Laramide orogeny (Tweto, 1975; Kluth and Nelson, 1988; Raynolds, 2002).

As the mountain blocks rose, erosion stripped some of the Cretaceous and older cover and shed synorogenic sediments into adjoining foreland basins, especially east of the Front Range (Tweto, 1975; Weimer and LeRoy, 1987; Raynolds, 2004). Coastal plain, swamp, and delta plain deposits constitute the subaerial facies of the Laramie Formation (Kl) that consists of sandstone, claystone, carbonaceous shale, and several economic coal seams in the lower several hundred feet; the Laramie is chiefly claystone in the upper several hundred feet (Roberts, 2005). The Laramie Formation was deposited over about the same time interval as the recessional shoreline deposits of the Fox Hills Sandstone (Kfh). An erosional unconformity marks the top of the Laramie-Fox Hills sequence in many areas, separating it from alluvial channel deposits (arkosic conglomerate and sandstone of the Arapahoe Formation) exposed east and south of this quadrangle (Weimer and LeRoy, 1987; Raynolds and Johnson, 2003).

Cobble conglomerate and arkosic sediment derived from eroded Proterozoic basement rocks mark the pulse of significant Laramide mountain uplift. These synorogenic sediments accumulated to the east in the Denver Basin (Arapahoe Formation and younger Tertiary deposits; Raynolds, 2004) almost continuously from about 70 Ma to 64 Ma (Raynolds and Johnson, 2003). Stacked, coarse alluvial-fan deposits and intervening carbonaceous siltstones in this sequence indicate that uplift was episodic. In the western and southern parts of the Denver Basin, sediment accumulation ceased after about 64 Ma and a deep-weathering profile (and paleosol) was formed prior to deposition of another coarse-grained alluvial fan after about 54 Ma (Raynolds and Johnson, 2003).

The Middle Park Formation and partly equivalent Coalmont Formation (₱c) are the synorogenic deposits preserved on the western flank of the Front Range uplift. They resemble the synorogenic deposits of the Denver Basin but seem to be significantly younger. The local basal unit of the Middle Park is the distinctive Windy Gap Volcanic Member (Kmw) (Izett, 1968) that consists of hundreds of feet of andesitic volcanic breccia, which originated from contemporary volcanic domes farther west in the Rabbit Ears Range (Izett, 1968). The Windy Gap was not directly dated but was tentatively assigned to the Upper Cretaceous? on the basis of Late Cretaceous pollen and spores in a single sample from overlying beds (Izett and others, 1963; Izett, 1968). All other samples of the Middle Park Formation contained Paleocene flora. In further contrast with the Denver Basin, both the Laramie-Fox Hills sequence and much of the upper Pierre Shale were eroded prior to deposition of the Middle Park and Coalmont Formations in this area (Izett, 1968; Tweto, 1975).

Most of the Middle Park Formation (₱Km, above the Windy Gap Volcanic Member) consists of a lower unit of coarse volcaniclastic conglomerate and sandstone that grades into an upper unit of arkosic sediment with abundant Proterozoic clasts (Kinney, 1970b). The transition in clast composition reflects the initial stripping of volcanic deposits off the uplifting mountain blocks, followed by erosion of the Proterozoic crystalline core. Local beds of coal are reported in the Middle Park Formation. Thickness of the upper part is highly variable owing to Laramide erosion prior to deposition as well as to time-transgressive onset of sedimentation. The formation is more than 6,000 ft thick in the Middle Park Basin to the west (Izett, 1968). The Coalmont Formation (₱c) is similar to the Middle Park and includes a lower unit rich in volcanic clasts and a thick upper unit that is more arkosic; total thickness may exceed 9,000 ft in the North Park basin (Hail, 1968). Pollen in the Coalmont Formation is exclusively Paleocene and Eocene (Hail, 1968).

Laramide Intrusive Rocks

The Laramide orogeny was accompanied by widespread alkalic intrusive activity in two main pulses between about 78–68 Ma and 62–45 Ma (fig. 6). These intrusions probably fed volcanic edifices on the rising mountain blocks, but concurrent and later erosion stripped most of those volcanic deposits (Tweto, 1975). The intrusive rocks are part of a broad, northeast-trending tract of early Tertiary igneous activity that became the locus of ore deposition known as the Colorado Mineral Belt (Lovering and Goddard, 1950; Tweto, 1975). The northern margin of the Colorado Mineral Belt in this quadrangle includes the mineralized intrusions at Jamestown, Ward and Sunset, and in the Audubon-Albion stock west of Ward (Lovering and Goddard, 1950; Pearson, 1980).

The principal magmas associated with the Laramide uplift are of two principal compositional types. Mafic-alkalic rocks include alkali gabbro, monzogabbro (Kmzb), monzonite, syenodiorite (Ksyd), monzodiorite (Kmzd), and shonkinite. Alkali-calcic rocks include syenite (₱sy), quartz monzonite (Kqm), and latite to rhyodacite (₱ql). Some intrusive centers were invaded by both types of magma over periods of 5 to 15 million years. The early intrusions (about 78–68 Ma) include both mafic-alkalic and alkali-calcic types. The younger intrusions (about 54–45 Ma) are mostly syenite and quartz syenite (₱sy) that post-date mineralization. Volumetrically minor sills intruded into Phanerozoic sedimentary rocks in the foothills-hogback belt at about 66–64 Ma are primarily rhyodacite and quartz latite porphyries (₱ql) (DeWitt, unpublished data; written commun., 2007).

Source reports (Gable and Madole, 1976; Gable, 1980a) describe many of the plutonic intrusive rocks as quartz monzonite or granodiorite, leading to the impression that calc-alkalic magmas were more widespread in this region during Laramide activity. Modern whole-rock chemical data show that mafic-alkalic magmas were the predominant types, and that most of the rocks are either alkali gabbro, syenodiorite, monzodiorite, or monzonite (DeWitt, written commun., 2007).

The Jamestown stock (fig. 6) consists of hornblende monzodiorite and minor hornblende-biotite quartz monzonite intruded at about 78–72 Ma (Gable, 1980a; Cunningham and others, 1994). A second intrusion near the end of Laramide activity consists of syenite and quartz syenite porphyry that was emplaced at about 56 Ma (Gable, 1980a; DeWitt, written commun., 2007).

Several intrusions in the Gold Hill-Sunset area (fig. 6) consist of leucocratic biotite syenite and biotite quartz syenite, including the Sugarloaf stock, Sunset stock, Burnt Mountain stock, Bald Mountain stock, and the Grassy Mountain stock. The Sunset stock yielded a K-Ar age of about 54 Ma (Gable, 1980a), whereas Sugarloaf Mountain and Bald Mountain appear to be distinctly younger (about 45 Ma) and post-mineralization (DeWitt, written commun., 2007).

The Audubon-Albion stock (fig. 6) is an irregular composite body of mafic-alkalic phases that are more and more siliceous eastward (Mathews, 1970). Alkali gabbro with small pods of pyroxenite on the west are intruded by syenodiorite and monzonite on the east. These rocks are intruded in turn by gray nepheline syenite and pink quartz syenite along the eastern margin of the composite stock. Hornblendes in the sequence range from 70 Ma to 65 Ma (progressively younger with increasing silica content), and biotite from the syenites indicate thermal closure at about 65 Ma (Gable and Madole, 1976; DeWitt, written commun., 2007).

The northern end of the Caribou stock (fig. 6) is exposed along the southern quadrangle boundary, south of the Audubon-Albion stock. Most rocks in this part of the Caribou stock are monzodiorite, associated with mafic pyroxenites. Hornblende ages are about 76 Ma (DeWitt, written commun., 2007).

An unusual Laramide dike intrudes the Pierre Shale (Kp) in the eastern part of Boulder near to the settlers' village of Valmont. This dike consists of augite shonkinite, a variety of nepheline-bearing alkali gabbro, similar to early mafic flows interbedded with the Denver Formation at Table Mountain near Golden, Colorado (about 25 km south; Weimer, 1996). The Valmont dike yielded a whole-rock Rb-Sr age of 64 Ma (Simmons and Hedge, 1978), also similar to the Table Mountain flows (Obradovich, 2002).

Several Laramide calc-alkaline magmas were intruded along the foothills-hogback belt just prior to uplift and tilting of the edge of the Front Range. These bodies form thick sills that intrude the Fountain (PPf) or Lyons (Pl) Formation just west of Boulder (Flagstaff Mountain sill), near the mouth of Left Hand Creek, and along the South St Vrain Creek just west of Lyons. These rocks are described as dacite (Braddock and others, 1988), but are chemically classified as quartz latite and rhyodacite (DeWitt, written commun., 2007). All of these sills yield K-Ar and fission-track ages between 66 Ma and 62 Ma (Hoblitt and Larson, 1975; corrected to revised decay constants).

Post-Laramide Sedimentary Rocks

Erosion characterized most of the Eocene Epoch and continued through much of the Oligocene in this region (Tweto, 1975; Epis and Chapin, 1975). Sedimentation resumed in the Granby valley area during the latest Oligocene with deposition of the basal Troublesome Formation (N₱t) in fault-bounded basins.

The base of the Troublesome Formation (N₱t) overlies an erosional unconformity that truncates folded beds of the Middle Park Formation and older units (Izett, 1974, 1975; Schroeder, 1995a). Initiation of deposition is locally well constrained by interbedded basalt flows (26.4 ± 0.9 Ma; Izett, 1974) and by interbedded laharic breccia derived from a latite plug (28.4 ± 2.3 Ma; Izett, 1974). The Troublesome consists of tuffaceous mudstone and sandstone with local conglomerate. Upper parts of the formation may be as young as 11 Ma (Izett, 1975).

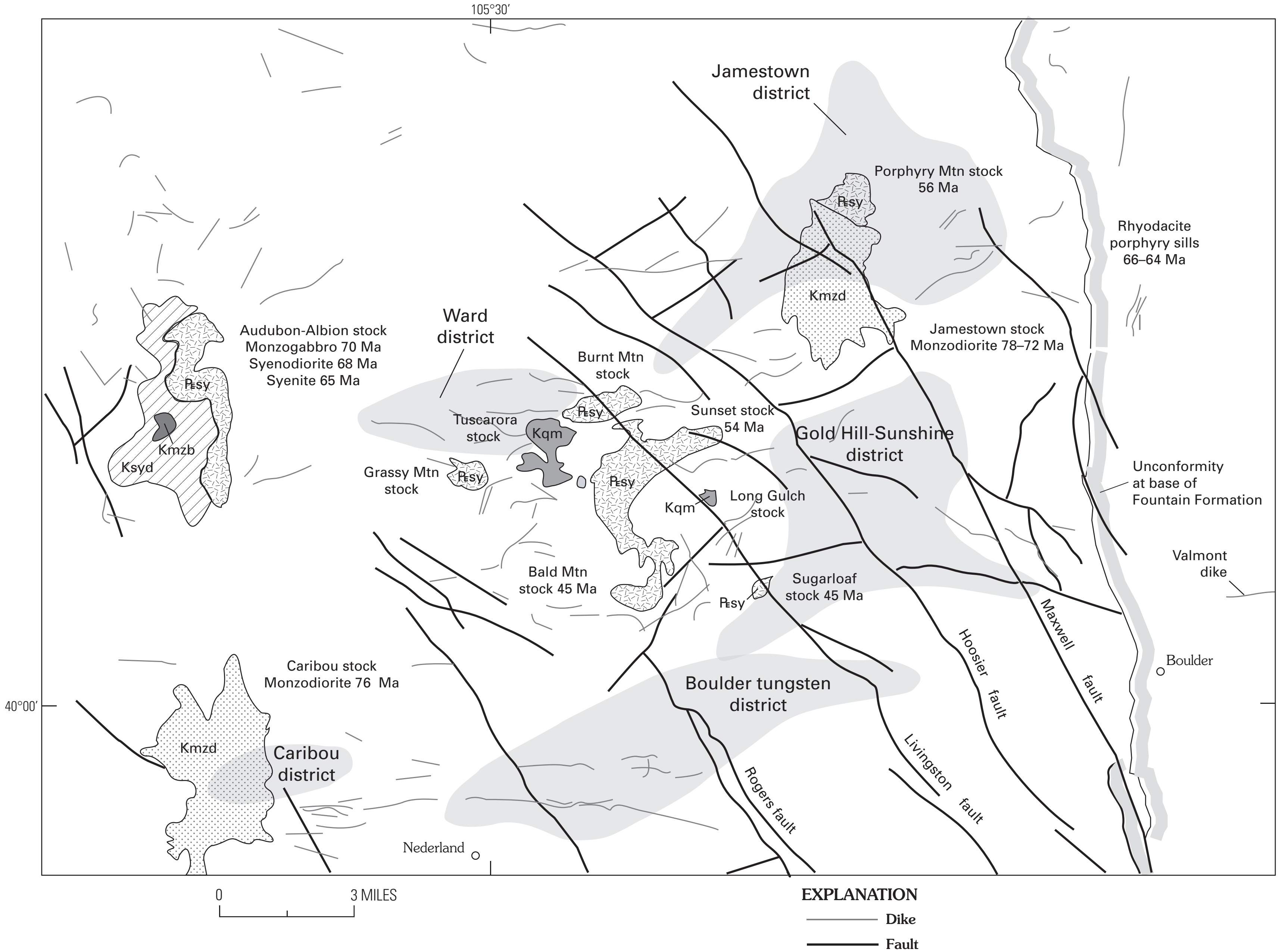


Figure 6. Sketch map showing Laramide intrusions in the Colorado Mineral Belt and mining districts in the Estes Park quadrangle.

Small areas of tuffaceous sandstone, mudstone, and conglomerate are present in the upper parts of the modern Colorado River drainage and underlie Oligocene volcanic rocks erupted from sources in the Never Summer Mountains. Most of these sparse remnants lie within paleovalleys, and some Proterozoic rock clasts in the conglomerate beds are as large as 3 ft in diameter (Braddock and Cole, 1990).

Post-Laramide Intrusive and Volcanic Rocks

Renewed magmatic activity is recorded throughout much of central and southern Colorado during the Oligocene Epoch, typified by intermediate calc-alkalic or differentiated high-silica magmas (Steven, 1975). The major centers of activity in the Estes Park quadrangle were the Braddock Peak intrusive-volcanic complex in the Never Summer Mountains and the Granby valley. The following descriptions are largely summarized from Izett (1974), O'Neill (1981), Braddock and Cole (1990), and Cole and others (2008).

The uplifted block of the Never Summer Mountains exposes the Braddock Peak intrusive-volcanic complex, including several intrusive bodies emplaced during late Oligocene, and surrounding areas that preserve some of the volcanic ejecta erupted from these magma chambers. Andesite porphyry (Ꝑap) at Tepee Mountain erupted in late Oligocene and produced andesite and basalt flows and laharic breccias that are preserved in a tilted fault block inclined eastward toward Little Yellowstone Pass, as well as in a paleovalley to the north.

The andesitic rocks are intruded by granodiorite and monzonite of the Mount Richthofen stock (Ꝑgdr), which was emplaced at 29.7 ± 3 Ma (Marvin and others, 1974). Farther south, the Mount Richthofen stock is intruded by granite and granite porphyry of the Mount Cumulus stock (Ꝑgc), which were emplaced at 28.2 ± 0.7 Ma (Marvin and others, 1974). Both of these intrusive centers are thought to have been the sources of rhyolitic welded tuff (Ꝑrw), and varied volcanic rocks that are preserved eastward in the tilted fault block in the Little Yellowstone area. Rhyolite ash-flow tuff and flow rock are preserved at Specimen Mountain, along the Continental Divide between Iron Mountain and Thunder Mountain, and at Seven Utes Mountain in the northwestern quadrangle corner. These rhyolites are geochemically similar to the granite of Mount Cumulus (Ꝑgc), are highly differentiated (strong europium anomalies), and are among the youngest volcanic rocks of the Braddock Peak intrusive-volcanic complex (Knox, 2005; Cole and others, 2008). They most likely erupted from the Mount Cumulus stock or adjacent areas marked by a pronounced negative gravity anomaly (see fig. 10, following; Cole and others, 2008).

The area surrounding Apiatan Mountain in the Trail Mountain quadrangle was also a center of intrusive and volcanic activity in late Oligocene time (Izett, 1974). Quartz latite porphyry (Ꝑqla), erupted at about 28.4 ± 2.3 Ma and produced laharic breccia (Ꝑqlx) that was interbedded with the lower Troublesome Formation (NꝐt). Rhyolite porphyry intruded in the vicinity of Porphyry Peaks and rhyolite flows are locally preserved. Basaltic and trachyandesitic flows, and a few intrusive plugs, are common in this area, and the flows are interlayered with the Troublesome Formation at several stratigraphic levels.

Fluvial Gravel

Scattered relics of bouldery, very poorly sorted, coarse-grained gravel, conglomerate, and sandstone deposits are preserved in several localities in the quadrangle. Some of these were investigated and reported by Madole (1982), and a regional map of similar or related deposits was compiled by Scott and Taylor (1986).

Wahlstrom (1947) examined some of these bouldery deposits (Ng) in the Niwot Ridge area west of Ward and concluded they were formed by glacial processes. He correlated these deposits, lying at altitudes as high as 12,300 ft, with similar bouldery deposits downvalley to the east at altitudes below 8,000 ft. Despite the similarity of these deposits to glacial till, Madole (1982) showed that a glacial origin was highly unlikely because the high-altitude gravels are higher than glacial ice is known to have amassed, and the low-altitude gravels are preserved far downvalley from the known limits of the younger, and more extensive, glacial deposits.

These bouldery gravels and conglomerates most probably were transported in paleovalleys that had developed during incision related to Miocene uplift (Scott and Taylor, 1986; Steven and others, 1997). One such paleodrainage (probable prior course of the Big Thompson River) is indicated by remnants of boulder gravel stranded on the mature rolling landscape north of Palisade Mountain (see fig. 12, following). Remnants of boulder gravel that cap ridges in the Bowen Mountain quadrangle at Gravel Mountain and between Blue Ridge and Apiatan Mountain appear to have formed in south-flowing drainage that pre-dates the modern Colorado River.

Quaternary Deposits

Quaternary deposits in the Estes Park quadrangle primarily reflect the climatic fluctuations of the last 1.8 million years related to the glacial-interglacial cycles. They comprise glacial deposits at high altitudes, stream deposits at lower altitudes beyond the mouths of mountain canyons, and various widespread thin deposits related to mass wasting and eolian processes.

Alluvial Deposits

The Quaternary geomorphic history of the Front Range reflects persistent erosion that has taken place since early Pliocene time. East of the mountain front, the South Platte River drainage cut through the Miocene sedimentary apron (Ogallala Formation) and exposed the tilted resistant layers in the hogback belt; upstream, river valleys became deeply entrenched in the mountain block (Steven and others, 1997). Progressive, probably episodic lowering of base-level on the piedmont has led to incomplete preservation of various surficial deposits in the eastern parts of the quadrangle. Similar processes have acted in the Colorado River drainage west of the Continental Divide. The main Quaternary fluvial units in the area are valley-floor alluvium and numerous terrace and pediment deposits that record fluvial and alluvial deposition beyond the mountain front (fig. 7).

Considerable study during the last five decades has been devoted to the fluvial-terrace and pediment deposits on the east side of the Front Range and their record of episodic drainage incision. Scott (1960, 1963a) described five prominent alluvial surfaces along the South Platte River drainage southwest of Denver, the geomorphology of the associated deposits and their soil and weathering characteristics, and established a nomenclature scheme. Subsequent work has largely validated these observations, allowed correlation of terrace and pediment deposits elsewhere along the mountain front (Colton, 1978), and provided better control on the ages of terrace and pediment formation (summarized by Madole, 1991).

Regional correlation of terrace and pediment deposits is challenging, particularly among older units, because numerical age criteria are few, relative dating methods are inexact, and local geology exerts considerable control on drainage evolution (Madole, 1991). For example, correlation based on terrace height above modern floodplain level can be unreliable, especially along the western margin of the Colorado Piedmont. Many older streams that flowed on gravel beds were pirated by younger streams that cut rapidly headward on erodable Pierre Shale (Kp). The disparity in stream level would be different in the non-pirated valley where the younger stream also flowed on a gravel-clad floodplain. Figures stated below for terrace heights represent average values in locations several miles east of the mountain front.

The oldest well-preserved deposit is the Rocky Flats Alluvium (Qrf). It includes an old alluvial fan that caps the Table Mountain mesa between Boulder and Lyons, as well as the elongate ridge (Table Top Mountain) north of Hygiene that marks the old course of the Little Thompson River during the early Pleistocene (Madole and others, 1998). The tops of these old gravel deposits are generally 260–300 feet above modern stream channels in this area. The Rocky Flats Alluvium is interpreted to be early Pleistocene based on its relation to younger, dated Verdos Alluvium (Qv).

Two small areas in the quadrangle preserve fluvial gravel at a higher position relative to the Rocky Flats terrace, and they are designated pre-Rocky Flats alluvium (Qprf). Their numerical age is unknown, but early Pleistocene is inferred (Scott, 1963a). These include gravel at the top of Gun Barrel Hill south of Niwot (at least 370 ft above Boulder Creek) and Haystack Mountain west of Niwot (360 ft above Left Hand Creek).

The next youngest fluvial deposit below the Rocky Flats is designated the Verdos Alluvium (Qv), which caps a few ridges along the Little Thompson River. Its upper surface lies about 170 ft above modern drainage channels, or about 100 ft below the Rocky Flats Alluvium (Qrf). The Verdos locally contains the Lava Creek B volcanic ash (639 ka; Izett and others, 1992; Lanphere and others, 2002) and is established as early middle Pleistocene (Madole, 1991).

The Slocum Alluvium (Qs) is widespread for several miles east of the hogback belt along the eastern margin of the Front Range. It forms a prominent pediment deposit in north Boulder, a terrace deposit along the Left Hand Creek drainage east to Niwot, and widespread pediment deposits on slopes surrounding Loveland. The Slocum terrace lies about 110–130 ft above modern drainage; its age may be as great as 240 ka, but it is not well defined (Madole, 1991).

The Louviers Alluvium (Qlv) forms a conspicuous terrace along reaches of Boulder Creek, Left Hand Creek, and St Vrain Creek, but it is largely absent from the Big Thompson drainage that had active glaciers in its headwaters. The top of the Louviers lies about 20–40 feet above modern streams. The Louviers Alluvium is interpreted to be lowland glaciofluvial debris derived from montane glacial deposits of Bull Lake age (Qtb) (roughly 200–130 ka), because of similarities in soil characteristics and clast weathering (Madole and Shroba, 1979; Madole, 1991). Much of the Louviers was probably deposited during late Bull Lake time.

The Broadway Alluvium (Qbr) forms a prominent terrace along Boulder Creek, South Boulder Creek, St Vrain Creek, and the Big Thompson River, 10 to 25 ft above modern stream beds. Characteristics of soil development and clast weathering are similar to glacial deposits of Pinedale age (Qtp). The Broadway is interpreted to have been deposited late in Pinedale time (30–12 ka; Madole and Shroba, 1979; Madole, 1991).

Alluvium in the modern drainageways is designated on this map as two distinct units based on geographic position in the mountain or plains regions. Piedmont streams east of the foothills-hogback belt locally preserve one or more low terraces, all of which are probably Holocene based on their position relative to Broadway Alluvium (Qbr). Mountain-valley alluvial deposits (Qva) are coarser and less structured due to steeper stream gradients and the availability of coarse side-stream detritus; they are probably both Holocene and late Pleistocene in age.

Granby Mesa lies between the confluence of the Colorado River and the Fraser River in the southwestern part of the quadrangle and is capped by coarse gravel and sand deposits (Qgv, Qgm, Qgo) that overlie eroded Tertiary and Mesozoic rocks (Schroeder, 1995a). Meierding (1977) correlated parts of these terrace deposits with the Pinedale, Bull Lake, and pre-Bull Lake glacial-interglacial cycles, based on their relative heights above the modern rivers. Geomorphic relations suggest Granby Mesa consists of a relatively old west-sloping alluvial-fan deposit (Qfo), a high terrace deposit (Qgo) about 220 feet above the Colorado River, an intermediate terrace deposit (Qgm) (Granby town site) at about 140 feet, and a low terrace deposit (Qgy) about 40 to 60 feet above both rivers. Other relics of terrace gravels

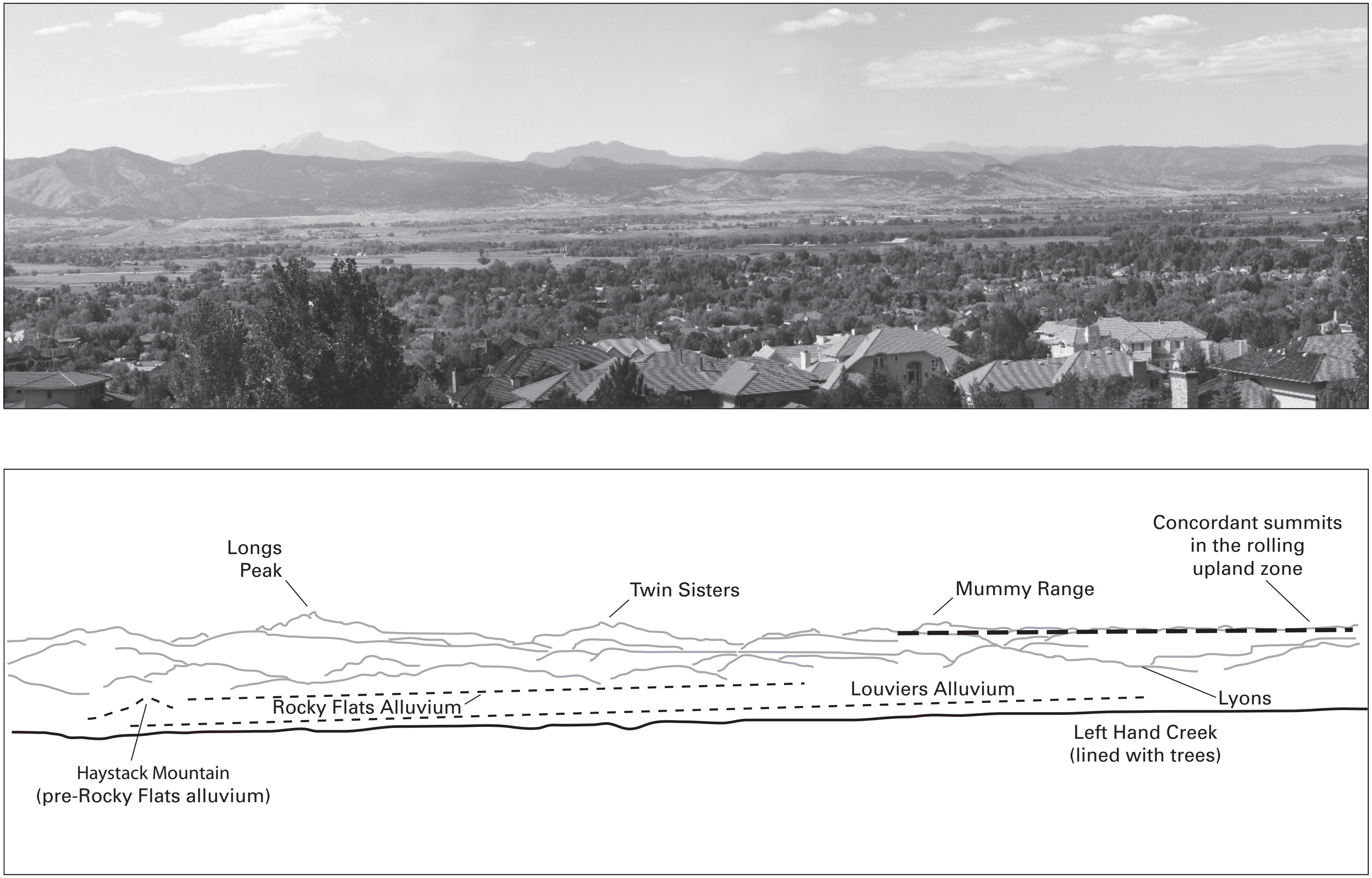


Figure 7. Panoramic photograph and drawing showing fluvial terraces, alluvial fans, and high-level erosion surfaces viewed northwestward from Gun Barrel Hill (J.C. Cole photograph, 2006).

preserved along the course of the Colorado River north and south of Lake Granby (Braddock and Cole, 1990) were also assigned relative ages based on elevation relations.

Diamicton

Several small areas within the quadrangle are underlain by poorly sorted bouldery deposits (Qd) that do not appear to be confined to paleochannels and thus differ from the fluvial gravel deposits described before. All lie at low positions in the modern landscape, which suggests they are relatively young, unlike the older fluvial gravels. These younger deposits are located in the Tahosa Valley east of Longs Peak. Richmond (1960) interpreted one of them to be pre-Bull Lake till (Qtpb), but a non-glacial origin seems more consistent with its location beyond the known limits of glacial ice. The southernmost deposit appears to be a remnant of a bouldery alluvial fan (Braddock and Cole, 1990) that formed on the eastern slope of Mount Meeker.

Glacial Deposits

The high mountain country of the Estes Park quadrangle records the history of Pleistocene and Holocene glaciation in considerable detail. The prominent glacial cirques along the Continental Divide, the classic U-shaped valley cross sections formed by glacial action, and the characteristic forms of lateral and terminal moraine ridges all attest to the extent of glaciation and are major scenic attractions in Rocky Mountain National Park (fig. 8).

Early work by Lee (1917, 1922) and Richmond (1960) identified the main elements of the glacial landscape here and established the evidence for multiple glacial advances and retreats. Subsequent detailed work by Madole (1969), Shroba (1977), Madole and Shroba (1979), and Birkeland and others (1987) has added further information about the glacial stratigraphy, soil characteristics, and ages of glacial events. Meierding (1977) mapped many of the glacial deposits on the west side of the quadrangle. A comprehensive regional compilation by Madole and others (1998) provides a succinct summary of much of the relevant information on Front Range glaciation. The following paragraphs are summarized from these sources.

Three principal ages of Pleistocene glacial activity are recognized in the Front Range. Pre-Bull Lake glacial deposits (Qtpb) chiefly consist of scattered remnants of deeply weathered till that predate late middle Pleistocene. Deposits of the Bull Lake glaciation (Qtb) consist of till and glaciofluvial sediment that formed during one or more advances and retreats in late middle and late Pleistocene time (roughly 300–120 ka). The Pinedale glaciation occurred during the late Pleistocene (about 30 ka to 12 ka) and produced the most extensive and best preserved moraines and glaciofluvial sediments (Qtp) in the Front Range.

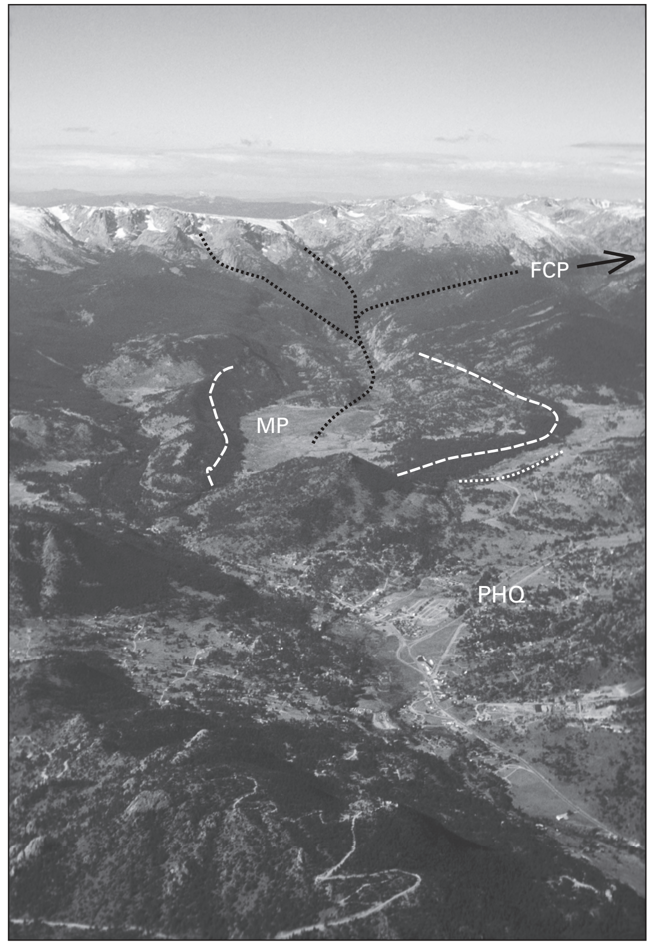


Figure 8. Oblique aerial photograph viewed westward over Moraine Park (MP) showing paths of major glacial streams (black dotted lines) that flowed from Forest Canyon Pass (FCP) and cirque basins along the Continental Divide (snow-capped skyline). South (left) and north (right) lateral moraines of the Pinedale glaciation are indicated by dashed white lines. Remnant of older Bull Lake lateral moraine is indicated by dotted white line west of Rocky Mountain National Park headquarters complex (PHQ) at middle right. Width of view is about 3 miles at PHQ (J.C. Cole photograph, 1974).

Glacial ice in this area was largely confined to valleys, and the intervening ridgelines appear to have remained exposed (ice-free) during most of the Pleistocene Epoch. Each of the three major glacial periods seems to have been similar in terms of ice accumulation; the lowest down-valley occurrence of undisputed till is at about 8,200 ft elevation, regardless of age.

Pre-Bull Lake till (Qtpb) is preserved along Green Ridge between Shadow Mountain Lake and Lake Granby. Small remnants are also identified at Harvey Island in Lake Granby and near the Granby Dam spillway. These deposits are all deeply weathered. Surface boulders on subdued moraines are strongly weathered and subsurface biotite-bearing boulders are disintegrated.

Till and gravel deposits of Bull Lake age (Qtb) are preserved beyond the Pinedale terminal moraines in almost all of the major montane valleys east and west of the Continental Divide. The ridge form of the moraine crests is typically subdued but still discernible, about a quarter of the moraine boulders are decomposed, and soils are developed to depths greater than 3 ft. Multiple moraine-crest ridges suggest multiple advances and retreats during the Bull Lake glaciation.

Till and glaciofluvial sediments of Pinedale age (Qtp) are widespread in the Estes Park quadrangle. Their youthful age is indicated by sharp moraine crests, thin soils, and fresh glacial boulders (some retain glacial polish and striations). Pinedale terminal moraines form the downstream impoundment for outwash sediments that cover flat-floored valleys (for example, Horseshoe Park, Moraine Park, and Kawuneechee Valley) that have only recently begun to incise. Kettles and kettle lakes are common in areas of Pinedale drift. Well preserved multiple moraines show that two or three advances and retreats occurred during the entire Pinedale glaciation; parallel lateral moraines and nested terminal moraines are conspicuous in many of the major glacial valleys.

Till in the southwestern part of the Estes Park quadrangle has not been studied in sufficient detail to systematically distinguish Pinedale-age from possible Bull Lake-age deposits (Pearson, 1980; Schroeder, 1995b). Most of this till appears to be Pinedale-age based on topographic characteristics and on work by Meierding (1977) around Lake Granby and drainages to the north. Till in the Arapaho Creek drainage east of Monarch Lake and south on Strawberry Bench is likely Pinedale-age because older deposits probably would have been disturbed and redistributed by the thick Pinedale glacial stream in this valley. Till deposits mapped in the west-central part of the quadrangle in the headwaters of Trail Creek below Gravel Mountain and west of Apiatan Mountain show sharp lateral moraine ridges that suggest Pinedale age, but till on the high-altitude slopes west of Porphyry Peaks shows less primary morphology and may be partly older.

Two additional minor glacial units are shown locally on the map. Rock-glacier deposits (Qr) are present in the cirque-basin areas along the Continental Divide and in the Never Summer Mountains at altitudes greater than about 10,000 ft. Most of these lobate, ice-cored masses of talus boulders have flowed downvalley during Holocene time, but some accumulation may date to late Pleistocene (Braddock and Cole, 1990). Organic-rich sedimentary deposits (Qo) (bog peat and carbonaceous silt) are shown in several areas underlain by glacial outwash and reworked, fine-grained glacial material. These deposits are preserved chiefly in the Roaring Fork valley east of Longs Peak, and west of Ward in the valleys of South St Vrain Creek and North Boulder Creek. Detailed records of late Pleistocene and Holocene glacial environments have been constructed from study of these lake and bog materials (Madole, 1976, 1980).

Other Surficial Deposits

Other surficial units of various origins are widespread in parts of the Estes Park quadrangle, but they have only been compiled in areas where they are especially thick, extensive, or important as indicators of geologic conditions or processes.

Colluvium (Qc) is widespread on moderate slopes adjacent to steeper terrain, for example in the foothills-hogback belt. Colluvium is also shown in high-altitude areas on the map where persistent frost action causes bedrock to fracture and move slowly downslope under the action of gravity.

Talus deposits (Qta) are largely confined to the bases of steep slopes in the high mountainous terrain. Talus cones and aprons are mapped most commonly at the toes of steep bedrock slopes that were scoured out by glacial ice.

Landslides are present in the Estes Park quadrangle in several contrasting geologic settings; some are old and seem to be relatively immobile at present while others remain active today. Several high-altitude landslide deposits (Qls) involve masses of Proterozoic crystalline rock, particularly south of Forest Canyon Pass and Trail Ridge Road on the west side of the Continental Divide. A large mass of biotite gneiss on the northwest-facing slope of Jackstraw Mountain has moved in the direction of dip of the gneissic foliation, and displays the typical hummocky topography of an active slide (Braddock and Cole, 1990; Raup, 1996). Numerous small, shovel-shaped slides have been identified within glacial till on the west side of the Kawuneechee Valley north of Grand Lake, and on the north shore of Lake Granby. The Troublesome Formation (N₽t) is also host to numerous small slides, probably due to its high content of silt and altered volcanic clays. Altered volcanic rocks at Porphyry Peaks have slumped westward in a large complex landslide mass that covers about 0.7 square mi. over Pierre Shale (Kp). Landslides are also common within the foothills-hogback belt. These masses involve loose jumbles of clay, silt, and sand with blocks as large as several meters that have moved down the dip slope of the Dakota Group (Kd), Fountain Formation (PIPf), and Lyons Sandstone (Pl), in order of decreasing frequency.

Some landslide masses within the Dakota Group (Kd) involve bedding-plane slip beneath the first sandstone member of the South Platte Formation that allowed large blocks to glide downdip and override eroded younger formations to the east. Some of these block-glide landslides are known to be Oligocene in age because they are overlain by Miocene Ogalalla Formation north of this quadrangle (Braddock, 1978). This distinctive class of landslides is described in greater detail in the following Structure section.

Eolian deposits (Qlo), consisting of sandy loess and silty loess, are common east of the foothills-hogback belt where they form blanketing sheets on low-relief surfaces. Thin loess (commonly 3–10 ft) is widespread and is extensively developed for agricultural use. These eolian deposits are inactive today; they accumulated downwind of alluvium and decomposed Cretaceous shale during high-wind conditions that accompanied late Pleistocene and Holocene glaciations.

Structure

The geology of the Estes Park quadrangle records a complex structural history spanning from the Paleoproterozoic to the late Cenozoic (see map). The earliest deformation is recorded by the metamorphic rocks of the Front Range, followed by structures that formed during intrusion of the Mesoproterozoic granitic rocks and subsequent faulting under deep-crustal conditions. Phanerozoic deformation related to the Ancestral Rocky Mountains orogeny is not explicitly recorded in the quadrangle, but is inferred from depositional patterns of sediments shed from the uplifted basement blocks. The Laramide orogeny, beginning in Late Cretaceous time, exerted a major influence on the Front Range region by bringing Proterozoic crystalline rocks up from depth to the same altitude as Late Cretaceous marine sediments. Subsequent uplift and block faulting, during and following volcanic activity in the western part of the quadrangle, has added structural complexity to the region and influenced the evolution of modern drainage. The most recent uplift of the central Front Range, in conjunction with climate change in the Pliocene, has produced most of the modern landscape and stripped much of the earlier Tertiary cover from the foothills and plains regions.

Paleoproterozoic Structures

Braddock (1970) estimated that more than 40,000 ft of marine sediment accumulated in this region south of the Wyoming Archean craton during the Paleoproterozoic. These sediments were chiefly marine shales and sandstones deposited by submarine turbidity currents from distant sources. Graded bedding and crossbedding preserved in areas of low metamorphic grade record the facing direction of the beds. These sedimentary structures, especially in the Drake quadrangle (Braddock, Nutalaya, and others, 1970) and areas to the north, define the axial trace of a major east-northeast-trending syncline that is overturned to the north. All beds dip southward, but those on the south limb of the fold are overturned. The syncline interpretation is further supported by the symmetrical outcrop of distinctive units of porphyroblastic biotite schist (Xbp) on either side of the syncline axis, and by the symmetrical position of the contact between knotted mica schist (Xbk) and quartzofeldspathic mica schist (Xbq). Nesse and Braddock (1989) found evidence for a parallel syncline north of the Estes Park quadrangle in a position that implies a fold wavelength greater than 9 mi; these first-generation folds can be traced for distances of more than 22 mi in outcrop.

The major first-generation folds appear to have formed during early diagenetic lithification while the sediments were still saturated with water. Braddock (1970) found that the axial zones of outcrop-scale first-generation folds showed evidence of shale injection across folded beds of sand. Layer-silicate minerals (detrital clays) were strongly aligned by mechanical sorting during this wet-sediment injection parallel to axial planes in the hinge zones of folds, whereas the same minerals remained parallel to overall bedding in the long limbs of these folds. Braddock (1970) and Braddock and Cole (1979) inferred that an early, low-grade metamorphic event overprinted these first-generation folds because metamorphic sericite and chlorite replaced the original detrital clay minerals and preserved the mechanical orientations formed during wet-sediment deformation. Nothing more is known of this earliest metamorphic event due to overprinting by the subsequent (dominant) regional metamorphism. The early event likely resulted from stratigraphic burial and structural loading during folding.

Second- and third-generation folds are widely developed in the area and they consist of antiforms and synforms with wavelengths of generally less than a mile (summarized from Braddock and Cole, 1979; Hutchinson and Braddock, 1987). The two fold sets are thought to be closely related in time and space because of their physical similarities. Second-generation folds (and the crenulation cleavages related to them) trend chiefly to the east or northeast; third-generation folds and cleavages trend northwest or north. Otherwise, both fold sets produce structures of similar scale and geometry, both produce axial-plane cleavage, both show steep fold axes, and both were active during intrusion of Boulder Creek Granodiorite (XgdB) or trondhjemite of Thompson Canyon (XjT) at about 1,715 ± 10 Ma. Mile-scale folds related to the second event are present in the Gold Hill quadrangle (Gable, 1980a) and west of the Continental Divide southeast of Granby (Pearson, 1980); second-generation cleavage and small folds are pervasive north of Highway 36 (Lyons to Estes Park). Third-generation folds are well developed along the Big Thompson River canyon in a northwest-trending zone near Drake, and related crenulation cleavages are widespread. This trend is known to be the younger of the two because it consistently overprints second-generation structures, although both fold trends probably represent essentially conjugate shear directions that formed during a protracted contractional event.

This deformation that produced the crossing sets of second- and third-generation folds may have been a continuation of the contraction implied by the major first-generation wet-sediment folding. The overturned syncline indicates contraction oriented broadly north-south, and the same orientation is suggested by the younger northeast and northwest (conjugate) folds.

All of this folding took place before the thermal peak of metamorphic recrystallization passed through these rocks at sillimanite-grade or lower conditions (summarized from Cole, 1977, 2004b). The index porphyroblastic minerals of biotite, garnet, cordierite, staurolite, and andalusite consistently overprint the crenulated fabric displayed by the matrix micas (sericite, chlorite, muscovite, and biotite). The porphyroblasts are irregular and their internal inclusions of quartz and feldspars show no helicitic patterns that would indicate rotation during growth. The matrix micas are shingled around the hinges of crenulation folds, rather than kinked by the folds, indicating the micas recrystallized after folding.

Higher-grade metamorphic rocks (above the first occurrence of sillimanite) show evidence of mineral growth and recrystallization more nearly coincident with peak deformation (Cole, 1977, 2004b). Both biotite and (especially) sillimanite are weakly to strongly aligned parallel to hinges of the late folds. Migmatite leucosomes form phacolithic masses in the hinge zones of late folds indicating anatectic melt was accumulating and migrating into fold cores during deformation (fig. 4).

The difference in timing between peak deformation and peak recrystallization in low- vs. high-grade metamorphic rocks is not contradictory. Rather, it is an expectable effect of the low thermal-conductivity of rocks and reflects the time-lag for transmission of heat through the deforming orogen. The deformation was probably broadly synchronous throughout the region, but the thermal maximum arrived later in the lower-grade terranes as it migrated upward and outward from the core of the orogen after folding had ceased.

Mesoproterozoic Structures

Mesoproterozoic time is chiefly marked in Colorado by emplacement of plutons and batholiths of the Silver Plume Granite and equivalents at about 1,400 Ma (Braddock and Cole, 1979). These granites are representative of widespread similar intrusions that have been identified across the North American continent between Labrador and southern California (Anderson, 1983). On the basis of similarities in composition, isotopic characteristics, and petrology, these extensive 1,400 Ma granites are interpreted as products of continent-wide melting of the lower crust, possibly related to fundamental changes in the underlying mantle at that time (Cole, 1977; Anderson, 1983; Anderson and Bender, 1989). This magmatic event cannot be sensibly related to any known or inferred contemporaneous oceanic arc-trench system nearby, and the lack of magmatic chemical diversity further distinguishes the 1,400 Ma granites from the magmatic suites typical of magmatic arcs (Anderson and Cullers, 1999).

The Mesoproterozoic granites have been described as “anorogenic” because most of the plutons are simple, elliptical structures that sharply crosscut the country rock and were emplaced in the absence of regional deformation (Anderson, 1983). In this part of the Colorado Front Range, the regional tectonic setting during emplacement is interpreted to be nearly neutral or slightly extensional, the latter inferred from the widespread swarm of north-northwest-trending mafic dikes that were intruded during the Silver Plume event (Cole, 1997; Anderson and Cullers, 1999). Contrary interpretations are discussed at the end of this section.

Longs Peak-St Vrain Batholith

The principal body of Silver Plume-type granite in the Estes Park quadrangle is the Longs Peak-St Vrain batholith (fig. 9; Boos and Boos, 1934; Cole, 1977), which extends from the eastern mountain front across the Continental Divide to the Colorado River valley. The batholith spans nearly 30 mi by 35 mi, is well exposed at elevations between 6,000 ft to over 14,000 ft, and emplacement-related structures are exceptionally well displayed in the alpine glaciated cirques along the Continental Divide. The following description of deformation related to emplacement of the batholith is summarized from Cole (1977).

The Longs Peak-St Vrain batholith was intruded into biotite gneisses at a time of neutral crustal stress, but the rising magmas profoundly deformed the surrounding rocks in the process. The simplest evidence of this is that compositional layering in the regional metamorphic rocks is characteristically steeply inclined in areas distant from the batholith (for example, the Big Thompson Canyon), but layering is sub-horizontal in the upper parts of the batholith along the Continental Divide (fig. 9). The rotation involved in this change of orientation occurred as the dry, viscous granite magmas rose plastically through the lower crust; the wall-rock gneisses also deformed plastically because they were nearly as hot and fluid as the intruding magmas. These structures related to emplacement of the Longs Peak-St Vrain batholith are designated fourth-generation folds in the regional scheme (Braddock and Cole, 1979). Small-scale folds of the fourth generation show crenulation of metamorphic biotite and sillimanite knots, as distinct from the pre-existing second- and third-generation folds. Recognizable older folds (second or third generation) are preserved between Estes Park and the Twin Sisters ridge area, but they were flattened and significantly rotated during emplacement of the granite.

The granite exploited the inherent weakness of the foliation direction in metamorphic rocks to intrude innumerable tabular sills, and the granite sheets and enclosing gneisses deformed and folded together, as shown by outcrop exposures and map-scale patterns. The trachytic flow-foliation within the granite defines the same patterns of folds as defined by compositional layering in the enclosing gneisses.

The external border of the Longs Peak-St Vrain batholith consists of several large lobate, bulbous domes separated by tight synformal screens of folded biotite gneiss (fig. 9). This pattern is particularly notable in the part of the batholith north of the Longs Peak area. These domes reflect gravity-driven deformation caused by the buoyant rise of the intruding magmas, and attendant sinking of the denser metamorphic wall rock. The tight synforms between the lobate domes reflect squeezing and crowding along the margins of the growing, rising domes. Broad, open swales and arches are evident in the shapes of tabular metamorphic-rock inclusions in the upper parts of the batholith, well displayed in the glaciated terrain. These structures also reflect the effects

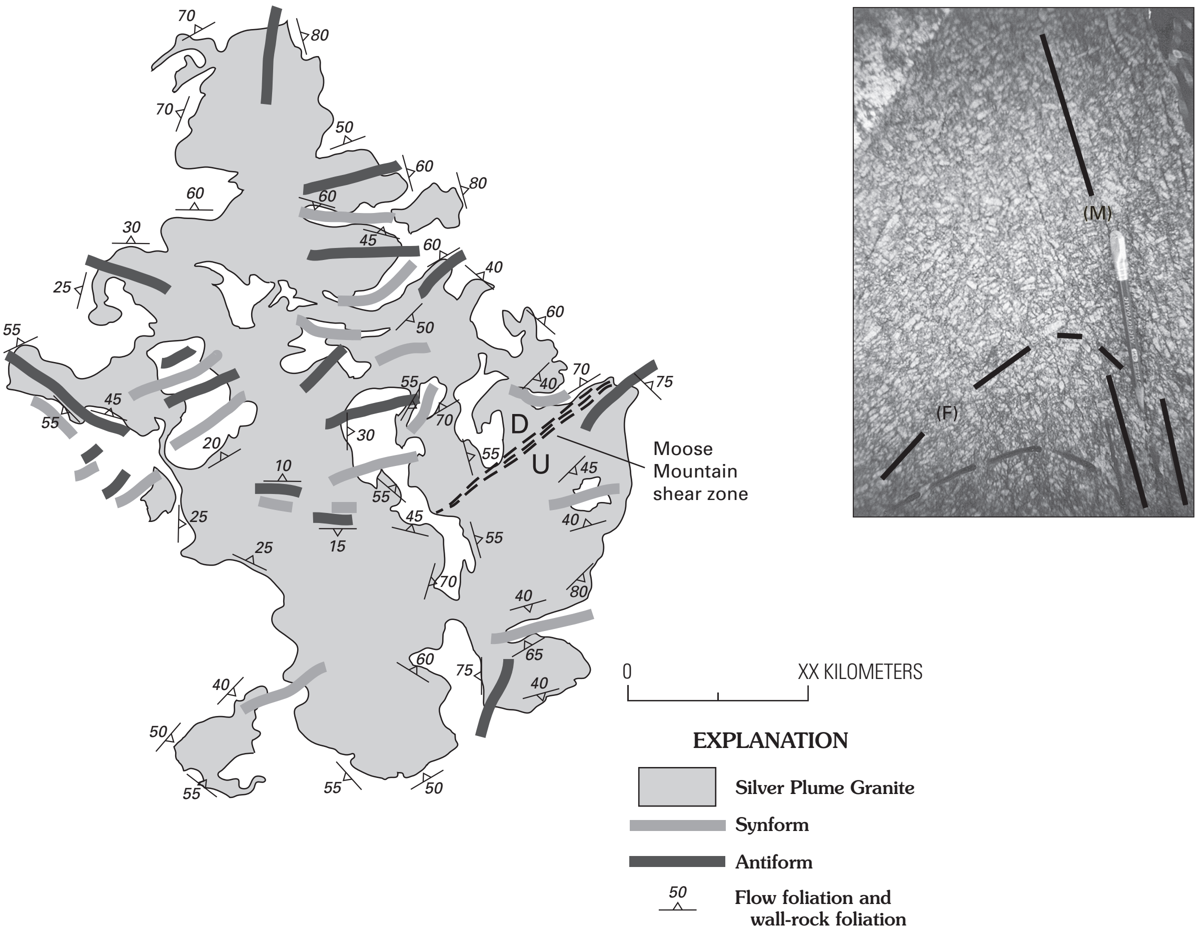


Figure 9. Map showing major structural elements of the Longs Peak-St Vrain batholith. Inset outcrop photograph shows detail within the Moose Mountain shear zone where magmatic flow-foliation of tabular feldspar phenocrysts (F) is re-oriented and overprinted by mylonitic shear planes (M). Pencil is about 4 in long.

of magma buoyancy relative to sinking country rock, not contraction or layer-parallel shortening.

Moose Mountain Shear Zone

The granite of Longs Peak (YgLP) and some of its local country rock are deformed in a broad zone of mylonite and protomylonite northwest of Lyons called the Moose Mountain shear zone (Braddock and Cole, 1979; Punongbayan and others, 1989). Cataclastic fabric within the zone trends east-northeast and dips steeply southward, and a steeply inclined streaking lineation is locally expressed by elongate clusters of minerals. The band of cataclasis varies in width from about 3,000 ft to 4,000 ft and the margins are gradational into undeformed rock. The Moose Mountain shear zone has been traced approximately 13 mi along strike from the eastern mountain front to the canyon of North St Vrain Creek east of Allens Park, where it feathers out and disappears. No structural zone of similar character was found west of Highway 7 along strike, nor was any mapped farther west (Gable and Madole, 1976; Pearson, 1980; Schroeder, 1995b).

The sense of offset across the Moose Mountain shear zone is inferred from regional and local relationships. Pelitic rocks north of the zone and east of Highway 36 are non-migmatitic (non-melted) and the zone truncates several amphibolite-grade mineral-zone boundaries (Punongbayan and others, 1989). All pelites south of the Moose Mountain zone are strongly migmatitic and must have melted at greater depth (higher pressure-temperature conditions), implying that the southern block has risen relative to the north. The steep mineral-streaking lineation within the mylonites is consistent with a reverse sense of displacement, as is the fact that map-unit contacts do not show major lateral (wrench) offset.

Cataclastic foliation along most of the trace of the Moose Mountain shear zone dips more than 60° S. Magmatic trachytic foliation within the granite of Longs Peak (YgLP) in this area has a similar northeasterly strike but chiefly dips northward. Abundant tabular microcline grains in the granite, which define the trachytic foliation, become progressively rounded as mylonite fabric becomes more pervasive in the rock, and feldspars are eventually rotated into parallelism with the enclosing cataclastic foliation. Selverstone and others (2000) cited various mineral textures and experimental deformation studies to conclude that the mylonite fabric probably developed while the granite was intruding. Cole (1977) also noted widespread slip surfaces farther west within the Longs Peak-St Vrain batholith that dislocate the trachytic foliation, aplite-filled “joints” within the granite, and some late dikes that locally show internal protomylonite fabric. He interpreted these features as local adjustments to volume reduction that occurred as a result of magma crystallization. The fact that the Moose Mountain shear zone feathers out to the southwest and disappears within the batholith is consistent with the notion that shearing ceased before magmatic crystallization was complete. In any event, the deformation in the Moose Mountain shear zone is older than the crosscutting Iron Dike (Ygb) (1,317 ± 50 Ma; Braddock and Peterman, 1989; Braddock and Cole, 1990).

Selverstone and others (1997, 2000) have interpreted the Moose Mountain shear zone as evidence of a regional contractional tectonic setting during the time of intrusion of Silver Plume-type granites. They point to the reverse slip-sense across the zone, the gradation between magmatic and cataclastic textures, and regional structural elements as evidence of north-northwest-directed compression during emplacement of the granite. However, the mafic dikes they site as indicators of extension direction are, first of all, older than the Longs Peak-St Vrain batholith and, second, mute about whether the regional compression axis was horizontal or vertical. Second, their interpretation of fourth-generation folds in and around the batholith as regional contractional structures ignores their diverse orientations (fig. 9) and abundant evidence that they resulted from gravity-driven vertical movements.

The simplest explanation of the Moose Mountain shear zone, in light of all the evidence for conditions during emplacement of the Longs Peak-St Vrain batholith, is that it formed during buoyant rise of the southern block (dominantly granite) relative to the northern block (dominantly metasedimentary rocks) while the magma was still partially molten. A similar interpretation is consistent with the facts in other sheared border zones of 1,400 Ma granites elsewhere in the region that have been proffered as examples of “regional compressional tectonics” (Selverstone and others, 2000, and included references).

Laramide Orogeny

Structures related to the Laramide orogeny are common in the Estes Park quadrangle. They include tilted basement blocks, drape folds, and reverse faults along the eastern mountain front, as well as thrusts and footwall folds on the western margin of the range (see map). Structures in these two areas are described separately below because they are kinematically distinct and because they differ substantially in age.

Eastern Margin of the Front Range

The most significant structure formed in the Laramide event lies covered beneath surficial deposits of the Colorado Piedmont on the eastern side of the quadrangle. The great asymmetric synclinal Denver Basin is the downwarped counterpart to the mountain uplift, and the Proterozoic basement rocks are depressed more than 13,000 ft along its axis (Higley and Cox, 2005). The deepest part of the basin coincides with that part of the mountain front where the Proterozoic crystalline rocks are thrust directly over the overturned Phanerozoic sedimentary rocks and the thrust-loading is greatest (Erslev, 1993). The basin is most asymmetric where the western edge is defined by such a thrust; the western limb is generally steep to overturned, whereas the eastern limb dips very gently westward (Haun, 1968). The thrust front is nearly continuous from Colorado Springs northward to about Coal Creek just south of the Estes Park quadrangle. The main thrust is blind northward through Boulder. The local “Boulder thrust” places steeply tilted Fountain Formation (PIPf) eastward over strongly overturned Cretaceous strata at the southern quadrangle boundary, but the fault cuts upsection in the hanging wall to the north and to the south, loses throw, and dies out within a few kilometers (Wrucke and Wilson, 1967; Wells, 1967).

From Boulder north to Left Hand Creek (about 12 km), the range front is marked by a simple homocline, moderate dips, and very limited folding in the Cretaceous and older strata. This style continues northward to Lyons, as the trend of the range front changes to north-northeast and the general inclination of beds declines to 20° or less. The axis of the Denver Basin also swings northeastward and migrates nearly 20 miles to the east beyond the quadrangle boundary (Haun, 1968).

At Lyons, and northward to the Wyoming State line, the range front steps progressively northeastward on northwest-trending faults (Braddock, Calvert, and others, 1970; Matthews, 1976; Braddock, Houston, and others, 1988; Braddock, Nutalaya, and others, 1988). These faults are east-dipping, high-angle reverse faults that raise the Proterozoic basement as much as 3,500 ft on the east (Denver Basin) side (see map). Sedimentary strata are primarily folded over the crests of these crystalline-block uplifts. Parallel anticlines and synclines east of the range front reflect reverse faults in the buried basement rocks. The west limb of the Denver Basin dips moderately overall in this region and it is folded along several of these blind, reverse-fault basement uplifts, but the overall basin is far more symmetrical and shallower at the latitude of Loveland than it is at Boulder (Haun, 1968; Erslev, 1993; Weimer, 1996).

Uplift of the eastern margin of the Front Range in the Estes Park quadrangle is inferred to be Late Cretaceous to Early Paleocene, based on evidence south of this area. The synorogenic sediments in the Denver Basin show that uplift began in Late Cretaceous time after about 70 Ma and continued more-or-less continuously until about 64 Ma (Raynolds and Johnson, 2003; Raynolds, 2003). This timing is consistent with the geologic relations in the Estes Park quadrangle.

Boulder-Weld Fault Zone

Upper Cretaceous rocks are broken by an anastomosing set of high-angle, northeast-trending faults in the southeast corner of the quadrangle known as the Boulder-Weld fault zone. These faults affect the upper Pierre Shale (Kp), the Fox Hills Sandstone (Kfh), and the overlying fluvial and coal-bearing strata of the Laramie Formation (Kl). The faults are well-documented through extensive subsurface mapping in coal mines and through surface mapping, and fault patterns have been inferred from widespread gas-well records (Spencer, 1961; Colton and Lowrie, 1973; Weimer, 1996; summarized in Roberts and others, 2001). This fault swarm displaces the Fox Hills Sandstone by as much as 500 ft along select strands (200 ft is more typical), but the net effect across several horst ridges and graben sags is essentially neutral; the overall southeasterly dip of strata is maintained across the Boulder-Weld fault zone from north to south, and the fault system does not significantly offset the Dakota Group (Kd) (Haun, 1968). The Boulder-Weld fault zone also appears to die out to the west and is not detected in the foothill-hogback zone adjacent to the Proterozoic core of the Front Range (Wells, 1967; Weimer, 1996).

Most of the faults in the Boulder-Weld zone were initially mapped and interpreted as north-dipping high-angle reverse faults that placed the Fox Hills (Kfh) above the Laramie Formation (Kl) (for example, Spencer, 1961). This interpretation is consistent with the synclinal bending of Laramie beds in the graben blocks and some arching of strata in the horst blocks, and with the presence of dip-slip slickensides on fault surfaces (Spencer, 1961). Kittleson (2004) also used extensive well records to demonstrate that the deformation is confined to the upper Pierre Shale (Kp) and younger units and is not seen in structure-contoured data for identifiable horizons deeper in the Pierre. He interpreted the Boulder-Weld fault zone to have formed by gravity-driven sliding along a surface of detachment within the upper Pierre Shale, causing imbrication of the section in the downdip direction toward the axis of the Denver Basin. In his interpretation, the Boulder-Weld fault zone is a consequence of the Late Cretaceous uplift of the Front Range and downwarping of the Denver Basin axis.

Weimer (1996, summary paper) interpreted structure-contour data quite differently and believes the Boulder-Weld fault zone is dominantly strike-slip. Tweto and Sims (1963) and Warner (1978) noted the apparent coincidence of a Proterozoic fault system (Idaho Springs-Ralston shear zone) with the Laramide Colorado Mineral Belt as indication of some persistent crustal zone of weakness. Weimer (1996) expanded on that observation and pointed out that the Boulder-Weld fault zone in the Denver Basin also aligns with these features in the mountain block, suggesting some connection among the three. The evidence for the Weimer strike-slip interpretation primarily rests on inferred differential offsets, or of thickness anomalies, in stratigraphic units ranging in age between 97 Ma and 62 Ma. His interpretation implies no fewer than eight distinct periods of wrench-fault displacement over about 35 million years, even though no fault dislocation appears in Lower Cretaceous or older strata. These apparent anomalies were interpreted from drill-hole records and generally amount to less than a few tens of feet over areas of many square miles. Kittleson (2004) noted that “correlations” of horizons within the Laramie-Fox Hills Formations, in particular, cannot be made with confidence because of inherent variations in thickness and facies formed during withdrawal of the Late Cretaceous Western Interior seaway.

We believe the data are insufficient to support the wrench-fault model of Weimer (1996), and prefer the interpretation by Kittleson (2004) that these stratigraphically and geographically localized faults arose from gravity-driven sliding within the upper Pierre Shale (Kp). Erslev (1993) similarly concluded this zone demonstrates “southeast-directed gravity detachments” in the Denver Basin. The absence of structural dislocation along the trend between the Boulder-Weld fault zone and the Idaho Springs-Ralston shear zone in the foothills-hogback belt is a strong indicator that the two structures arose independent of each other, regardless of coincidental location and trend.

Western Margin of the Front Range

The Never Summer thrust (O’Neill, 1981) marks the western structural margin of the Front Range uplift. This thrust places broken Proterozoic rocks westward over Upper Cretaceous Pierre Shale (Kp) and Paleocene and Eocene Coalmont Formation (Ꝑc) (partial lateral equivalent of the Middle Park Formation), which are deformed in west-vergent overturned footwall folds (see map). The Never Summer thrust is particularly well exposed because it was domed upward by intrusion of the subvolcanic stocks during the Oligocene and now lies thousands of feet higher than the level at which it was emplaced. The structural overhang of Proterozoic rocks on this thrust may be about 6 mi, based on its exposed position east and west of the Never Summer Mountains. The thrust can be traced southward through the Bowen Mountain quadrangle (Braddock and Cole, 1990; Braddock, unpub. mapping, 1985), but it is obscured by Oligocene volcanic and intrusive rocks near Apiatan Mountain (Ꝑɛqla, Ꝑɛr) (Izett, 1974) and by Neogene deposits of the Granby basin (Schroeder, 1995a). Westward-vergent folding is displayed by the Windy Gap Volcanic Member (Kmw) of the Paleocene and Upper Cretaceous Middle Park Formation in the southwestern corner of the quadrangle.

Erslev (1993) and Erslev and Selvig (1997) summarize their interpretation that the Never Summer thrust is a back-thrust that is secondary to a dominant east-vergent structure that underlies the Front Range uplift. However, uplift of the two flanks of the range are distinct in age and may not be directly related. The Never Summer thrust overrides Paleocene beds of the synorogenic Coalmont Formation (₱c) and probably late Paleocene beds of the Middle Park Formation (₱Km). These formations rest on eroded Pierre Shale (not on the Laramie-Fox Hills couplet), indicating that erosion was continuing in Late Cretaceous-early Paleocene time while synorogenic sediments were accumulating in the Denver Basin (Raynolds, 2003). Therefore, the Never Summer thrust may have formed under different stress conditions operating perhaps 10 million years later than initial uplift of the eastern margin of the Front Range.

Sandstone Dikes

Pearson (1980) reported lenticular and dike-like bodies of cemented, quartzose sandstone within the Proterozoic crystalline rocks along the trace of the west-northwest-trending Arapahoe Pass fault zone in the south-central part of the Estes Park quadrangle (see map). Schroeder (1995b) reported similar bodies in a north-trending fault zone farther west in the Strawberry Lake quadrangle. The following description is summarized from Pearson (1980), and from a more recent assessment by Dockal (2005).

The sandstone dikes are exposed for a distance of about 2.5 miles along the Arapahoe Pass fault zone and over a vertical elevation range between about 10,800 ft and 11,900 ft. They form discontinuous bodies that are elongate along the trend of the Arapahoe Pass fault and range in width from a foot or so to about 60 ft. The rock is internally brecciated, the Proterozoic wall rock is sheared, and all rocks are mineralized by secondary hematite, silica, and dispersed lead, zinc, and silver minerals. The sandstone within the dikes is mostly white (secondary tan, gray, or reddish-brown coloration) and consists almost exclusively of well-rounded quartz grains with a few percent feldspar. The dikes contain angular fragments of granitic wall rock, feldspar, and mica.

Similar sandstone dikes were described southwest of Denver by Scott (1963b) and a regional summary and petrographic study was reported by Harms (1965). Controversy remains regarding the origin and age of these distinctive structures. Both Scott (1963b) and Harms (1965) concluded that the sand most likely infiltrated fractures in the Proterozoic basement as a granular slurry from sources at higher structural levels. The uniformity and maturity of the rounded sand grains and heavy-mineral suites suggested the most likely source was the Upper Cambrian Sawatch Quartzite (Scott, 1963b). However, the Sawatch was removed from most of this area by erosion during the late Paleozoic Ancestral Rocky Mountains orogeny. Harms (1965) pointed to the close correspondence between the trend and location of sandstone dikes and the (flexed) leading edge of the main east-vergent uplift of the Rocky Mountain front between Colorado Springs and south Denver, indicating a Laramide age for the structures. He also noted that red-brown siltstone chips were entrained in the sandstone slurry in several locations and that such materials were only common in the Fountain Formation (ℙPf) in this region. Thus, the source of the sand may be higher in the Phanerozoic section than the Pennsylvanian.

The sandstone dikes in the Arapahoe Pass fault zone and west in the Strawberry Lake quadrangle offer no additional clues to the origin of these features. Their position on the landscape is nonetheless strong evidence that open fractures extended deep into the Proterozoic crystalline basement at time(s) in the past, and are indications of localized, significant high-level extension (fracturing) within the uplifted Front Range basement block. We believe that the preponderance of evidence favors the Laramide age for emplacement. The source of the sand could be the mature eolian quartzites of the Lyons Sandstone (Pl), which were loosely cemented with calcite in the principal area of the sandstone dikes.

Post-Laramide Structures

The deformation history of the Front Range area following the Laramide orogeny is mostly contained in deposits preserved outside the Estes Park quadrangle. The Laramide orogeny concluded in the area of the Estes Park quadrangle with emplacement of the Never Summer thrust. The underlying beds of Middle Park and Coalmont Formations were regionally folded and locally overturned in the thrust footwall.

The late Eocene Epoch was marked by reduced uplift; accelerated weathering due to intensely warm, moist climate; and widespread erosion throughout the Southern Rocky Mountain region (Epis and Chapin, 1975). The result of these factors was the formation of a topographically subdued landscape over large areas characterized by rolling hills and broad valleys (Epis and Chapin, 1975; Steven and others, 1997) between relics of mountainous terrain. This late Eocene eroded landscape is well preserved southwest of Denver beneath ash flows of the Wall Mountain Tuff (35 Ma), but no marker deposits of this age are present in the northern Front Range. Scott and Taylor (1986) inferred that much of the east-central part of the Front Range block in the Estes Park quadrangle, lying between about 8,000 ft and 9,000 ft elevation, represents relics of the Eocene landscape, but it is equally or more likely that this area of broadly concordant hilltop summits formed during Oligocene-early Miocene time (Evanoff, 1990; Steven and others, 1997).

Oligocene sediments and volcanic strata are only present in a few areas of the Estes Park quadrangle, most having been removed by subsequent erosion. The upper Oligocene and lower Miocene Troublesome Formation (N₱t) consists of tuffaceous siltstone, sandstone, and conglomerate that were deposited unconformably over Middle Park Formation in the Granby area. The Troublesome appears to have filled basins and half-graben that were bounded by north-south or northwest-trending high-angle faults.

Braddock Peak Intrusive-Volcanic Complex

Oligocene deposits exposed in the northern Never Summer Mountains record some of the history of the Braddock Peak intrusive-volcanic complex (O'Neill, 1981; Braddock and Cole, 1990; Cole and others, 2008). The granodiorite of the Mount Richthofen stock (₽gdr) (29.7 ± 3 Ma) and the granite of the Mount Cumulus stock (₽gc) (28.2 ± 0.7 Ma) are the intrusive roots of volcanoes that erupted here in the late Oligocene. Thick deposits of rhyolite ash, welded ash-flow tuff, and rhyolite flows are preserved at Specimen Mountain, Thunder Mountain, Seven Utes Mountain, and upland areas north of the quadrangle (fig. 10). Interbedded volcaniclastic sediments and possible cooling breaks in the tuff sequences attest to a considerable span of time marked by variable volcanic activity. Somewhat older andesitic flows and breccias are interbedded with rhyolite flows, volcaniclastic sediments, and ash-flow tuffs that are preserved in a fault-bounded half-graben in the Little Yellowstone area (Colorado River headwaters). Tilting and abnormal thickness in this and other areas of the Braddock Peak complex indicate extensional faulting accompanied volcanic activity.

Emplacement of the stocks in the center of the range caused deformation of the Never Summer thrust. The thrust is generally flat-lying or gently inclined to the east, except where it was bowed up across the apex of the Mount Richthofen stock (Corbett, 1968; O'Neill, 1981; Braddock and Cole, 1990). Between Mount Cindy and Bearpaws Peaks, the thrust is gently inclined to the west where it cuts upsection through Coalmont Formation in the footwall (O'Neill, 1981), indicating post-thrust tilting.

The rhyolitic ash-flow tuffs and flows at Thunder Mountain and areas north comprise two major eruptive deposits and possibly more. The total thickness of rhyolite locally exceeds 1,000 ft but declines notably toward the north (Cole and others, 2008).

The rhyolitic ash-flow tuffs and flows provide interesting information about the timing and amount of faulting that took place during eruption. The lower unit is a crystal-rich ash-flow tuff that contains conspicuous phenocrysts of euhedral, brown quartz. It is many hundreds of feet thick in tilted sections at Little Yellowstone, Seven Utes Mountain, and below Thunder Mountain, as well as at Gould Mountain just north of the quadrangle (fig. 10). From Thunder Mountain north, this ash-flow tuff thins rapidly due to erosion and is overlain by subhorizontal rhyolite flows that also thin northward (Cole and others, 2008). The correlation of unusually thick sections with tilted beds suggests that normal faulting occurred during eruption. Local lenses of very coarse breccia that contain clasts (as large as 30 ft in diameter) of various volcanic rocks, Proterozoic gneiss, and Phanerozoic sediment are interbedded with rhyolite welded tuff at Seven Utes Mountain and Little Yellowstone (Lee, 1917; O'Neill, 1981). The angular nature of these breccia deposits is consistent with rapid deposition possibly due to local fault displacement.

Both of the rhyolite units are highly silicic and differentiated (Knox, 2005) and most similar in composition to the Mount Cumulus stock. The ages of eruptive and intrusive rocks are similar within analytic error (Knox, 2005). The Mount Cumulus stock would seem to be a logical source for the extrusive rhyolites but no caldera structure is apparent in its vicinity. Gravity data show a conspicuous oval low anomaly that includes the Mount Cumulus stock and areas to the west underlain by a blind granite porphyry (proved by drilling; Pearson and others, 1981), a tuff-breccia ring intrusion (Metzger, 1974), and thermally metamorphosed Coalmont Formation. Taken together, this evidence indicates a more widespread, low-density granite body at shallow depth that probably fed the rhyolite eruptions (Cole and others, 2008).

Colorado River Fault Zone

The upper Colorado River drainage north of Grand Lake is a conspicuously linear feature that trends nearly due north for a distance of about 17 miles to the site of abandoned Lulu City. The glaciated floor of the Kawuneechee Valley exposes widespread brecciation, silicification, and iron-oxide mineralization in shattered Proterozoic rocks that indicate a substantial brittle fault zone (Braddock and Cole, 1990). The fault zone has only been mapped in the lower parts of the valley and it does not continue northward through the Oligocene volcanic rocks of the Braddock Peak intrusive-volcanic complex that cap Thunder Mountain, Lulu Mountain, and Iron Mountain (north of this quadrangle). Rather, the fault zone appears to turn northwestward and splay into smaller breccia zones that feather out. Some of the northwest splays offset granodiorite of the Mount Richthofen stock (₽gdr), which would indicate activity younger than 29.7 Ma if the faults are connected. The fault zone may be displaced by younger movement on the northwest-trending fault zone that traverses Specimen Mountain, the Little Yellowstone area, Thunder Pass, and the headwaters of the Michigan River.

The Colorado River fault zone also lies on strike with the Laramie River valley to the north, and Chapin (1983) has speculated on regional considerations that this long, combined lineament might have been a Laramide-age wrench fault zone. No local evidence supports this speculation. The Laramie River valley is controlled by Laramide-age reverse faults that place Proterozoic rocks in the Front Range and Medicine Bow Mountains over synclinal Mesozoic rocks in this valley (Beckwith, 1942). The northwest-trending, Mesoproterozoic Iron Dike (Ygb) does not appear to be offset across the Laramie River valley (Pearson and others, 1982).

Other north-trending faults are present in the Never Summer Mountains and all are mapped as normal-offset structures (O'Neill, 1981). They primarily displace the Oligocene intrusive rocks in the center of the range upward relative to the surrounding country rocks; it is possible the faults reflect late adjustments to the emplacement of the low-density stocks during volcanic activity.

Block-Glide Landslides

The east-dipping strata of the foothills-hogback belt have been substantially disrupted by large landslide complexes, exclusively within strata of the Dakota Group (Kd). Detailed mapping by Braddock and Eicher (1962) and Braddock (1978) showed that these landslides are characterized by gravity-driven detachment of parts of the Dakota section along stratigraphic zones of weakness, causing the detached block to slide over the downslope strata and override the contemporaneous land surface. The detached block was folded in the process of sliding over ridges in the underlying, intact strata. The block-glide landslide complex north of Boulder involves only the first sandstone member of the South Platte Formation, but most others farther north to Carter Lake Reservoir involve nearly the entire Dakota Group (Braddock, 1978; Braddock, Nutalaya, and others, 1988).

Similar landslide complexes have been mapped as far north as the Wyoming State line (Braddock and Eicher, 1962; Courtright and Braddock, 1989). Evidence there indicates that some of the block-glide landslides formed during Oligocene time because the slide complex locally overlies basal conglomerate of the White River Group and is overlain by Miocene Ogallala Formation (Braddock, 1978). On geomorphic grounds, it appears that many other block-glide landslides formed during the Pleistocene because the elevation at which the detached block broke and overrode the land surface coincides with one of the major pediment surfaces that formed during early Pleistocene erosion. Braddock (1978) concluded that the landslides were triggered by erosional removal of unconsolidated material at the east base of the hogback slope, probably during prolonged wet periods.

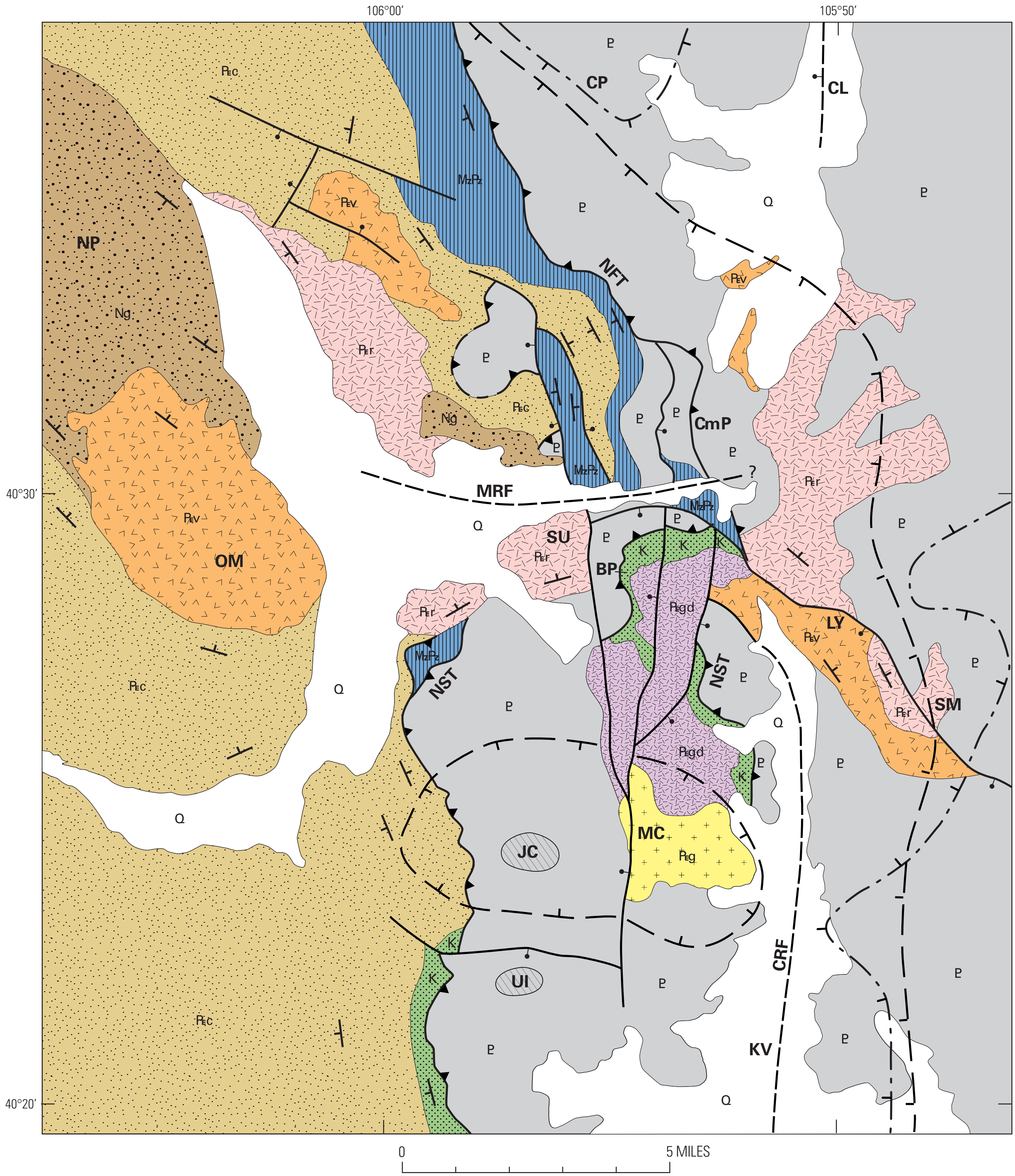
Late Miocene–Pliocene Uplift and River Incision

Following the late Oligocene and early Miocene volcanic eruptions from the Braddock Peak intrusive-volcanic complex in the Never Summer Mountains, the Miocene Epoch appears to have been largely a time of continued erosion and landscape smoothing across the Front Range (Scott, 1975; Steven and others, 1997). No deposits of early Miocene age are preserved in the mountains or on the eastern flank of the uplift (Courtright and Braddock, 1989; Steven and others, 1997). West of the Front Range, siltstone, sandstone, and conglomerate of the Troublesome Formation (NPt) accumulated in the Granby basin, starting in the late Oligocene and continuing into late Miocene (to about 11 Ma; Izett, 1974, 1975).

Middle Miocene time marked the onset of significant renewed uplift of the Front Range in this area (Steven and others, 1997; Bolyard, 1997). Eaton (1986, 1987) assembled regional evidence to argue that this younger uplift established the main topographic elements of the modern Southern Rocky Mountains province between Casper, Wyo. on the north and El Paso, Tex. on the south (fig. 11). The uplift is subcontinental in scale and most likely reflects thermal rise of a welt of aesthenospheric mantle beneath the region and related lithospheric thinning (Alvarado Ridge of Eaton, 1986). The axis of this broad ridge is marked by the location of the Rio Grande rift, a north-south trending series of normal faults and deep local basins that formed concurrent with (and directly resulting from) the regional uplift (Eaton, 1986; Chapin and Cather, 1994).

In the area of the Estes Park quadrangle, this uplift is suggested by the transport of coarse boulder gravel deposits (Ng) in east- and northeast-trending paleocanyons of the early Big Thompson River and South St Vrain Creek canyons (Scott and Taylor, 1986), and late Oligocene and Miocene southwest-flowing drainage of the early Colorado River system (Izett, 1968, 1974; Braddock, unpub. mapping, 1985). These gravel deposits that mark the positions of old canyons define fairly straight drainages that suggest the stream gradients were moderate to steep across the Proterozoic crystalline rocks. Steven and others (1997) noted that these Miocene drainages cross the pre-existing Oligocene drainages of the White River Group at a high angle and indicate regional post-Oligocene tilting of the Front Range block toward the northeast (that is, uplift toward the southwest). Miocene sand, gravel, and conglomerate of the Ogallala Formation (about 17 Ma to 6 Ma; Izett, 1975) record the middle to late Miocene uplift and erosion of the modern Front Range and Laramie Range. The fact that the Oligocene White River Group is eroded south of Fort Collins but the Ogallala persists southward is consistent with northeastward tilting of the range margin (Steven and others, 1997) during Ogallala time.

The change in general tilt direction appears to be somewhat more complex than described above, based on geomorphic relations in the Estes Park quadrangle and farther north. The Miocene paleovalleys in the northern Front Range are depicted (Scott and Taylor, 1986) as discharging directly to the east or northeast across the foothills-hogback belt. However, the paleovalley of the Big Thompson River apparently did not continue straight northeastward from the high-level gravel remnants north of Palisade Mountain because a continuous ridge of Proterozoic rock stands at higher elevation than those gravel deposits (fig. 12; Braddock, Nutalaya, and others, 1970). Rather, it appears that the Big Thompson drainage was deflected southeastward toward the modern canyon mouth, as indicated by the easternmost and lowest boulder gravel deposit preserved on the side of the modern Cedar Creek valley. Very minor faults in the valley floor seem insufficient for major post-Miocene uplift of the Green Ridge block, and so the Green Ridge prominence probably existed during the Miocene Big Thompson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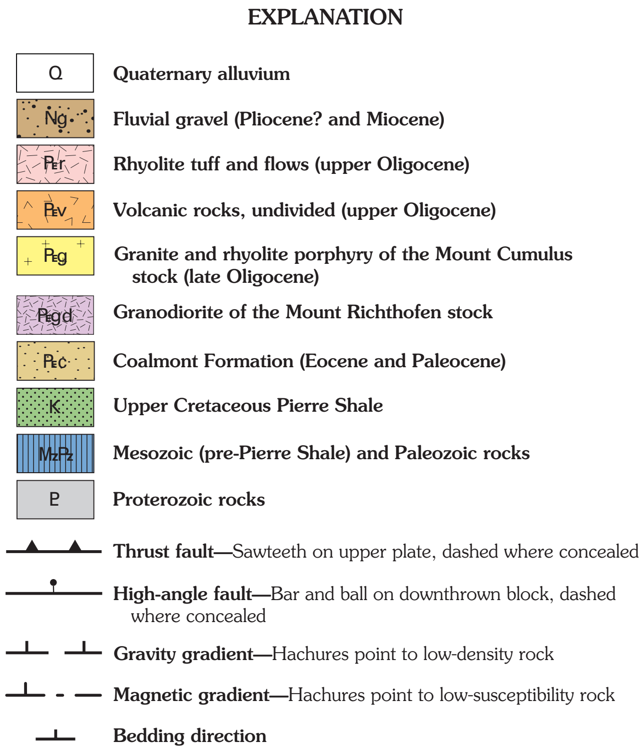

Figure 10 (above and previous page). Structural sketch map of Oligocene-Miocene volcanic and tectonic features related to the Braddock Peak intrusive-volcanic complex in the northern Never Summer Mountains. Geologic features shown: oval gravity low (dashed, hachured line enclosing buried granite porphyry at Jack Creek (JC) and Mount Cumulus (MC) stock); tuff-breccia ring in Upper Illinois River (UI) drainage; Colorado River fault (CRF); Michigan River fault (MRF); North Fork thrust (NFT); Never Summer thrust (NST). Geographic features shown: Braddock Peak (BP); Chambers Lake (CL); Cameron Pass (CmP); Clark Peak (CP); Kawuneechee Valley (KV); Little Yellowstone (LY); North Park (NP); Owl Mountain (OM); Specimen Mountain (SM); Seven Utes Mountain (SU).

paleodrainage. It was probably never covered by the Ogallala Formation deposits. Similar relationships are evident along the old course of Buckhorn Creek just north of this area (Braddock and others, 1989).

Deposition of the Ogallala Formation continued until latest Miocene time and appears to have led to fluvial aggradation west of the hogback belt in this area. The surface of terminal Ogallala deposition must have had a gentler slope than the middle Miocene streams that flowed in the fairly straight paleocanyons. This conclusion is deduced from the present meandering courses of the canyons of the lower Big Thompson River, Little Thompson River, North and South St Vrain Creeks, and Boulder Creek (Steven and others, 1997; Cole, 2004a). These streams today flow through steep-sided canyons cut in granitic and metamorphic rocks, where stream courses show little regard for structural weaknesses in the basement. The meanders in these river courses indicate they were established in low-gradient drainages on top of the Ogallala Formation, probably with broad floodplains, and then subsequently superimposed on the Proterozoic crystalline rocks during later uplift (Bolyard, 1997; Steven and others, 1997; Cole, 2004a). The fact that entrenched meanders are preserved several miles west of the foothills-hogback belt suggests that the Ogallala cover probably extended at least that far west as well.

Erosion has dominated Front Range geomorphology since the end of Ogallala Formation deposition in late Miocene (Steven and others, 1997; Leonard, 2002). The main drainages tied to the South Platte River have excavated almost all of the Oligocene and Miocene deposits east of the Continental Divide in this area, and the prominent hogback ridges of the foothills belt were exhumed from overlying Cenozoic cover (Steven and others, 1997). This erosion was certainly triggered in part by renewed uplift of the Front Range block. Cross-sectional forms of present canyons through the Proterozoic basement rocks show that the most recent deepening has occurred rapidly, leading to very steep inner-canyon walls (Bolyard, 1997; Steven and others, 1997). The entrenched meanders show that the rapid incision (post-5 Ma) did not allow time for streams to adjust to contrasts in canyon-wall fabric or rock types.

The accelerated erosion during Pliocene time may also reflect the marked change to cooler, wetter climate following the Miocene (Molnar and England, 1990; Crowley and North, 1991; Krantz, 1991; Bluemle and others, 2001). However, most studies conclude that climate-induced accelerated stream power alone is insufficient to explain the incision. McMillan and others (2002) examined the Ogallala Formation along the Colorado-Wyoming State line and concluded that the eastward tilt of beds is greater than it could have been during deposition, implying both tectonic uplift and isostatic uplift related to removal of overburden from the range block. Leonard (2002) examined the Pliocene and Quaternary incision of the North and South Platte River drainages and concluded, as well, that both tectonic uplift and isostatic adjustment to erosion were required to account for the gradients of modern streams and post-depositional warping of the Ogallala.

Pliocene and younger faults that might have accommodated this tectonic uplift of the Front Range core have not been positively identified. Steven and others (1997) suggest that the Pliocene uplift of the Southern Rocky Mountains involved diffuse adjustments in irregular fault-bounded blocks within the main ranges. One fault that may have accommodated some of this late uplift strikes northwest through Sunshine, Jamestown, and Allens Park, along the eastern base of Longs Peak, and continues over the Continental Divide at the head of Forest Canyon. North-trending fault zones in the upper Colorado River valley are also likely candidates for Pliocene-Quaternary displacement.

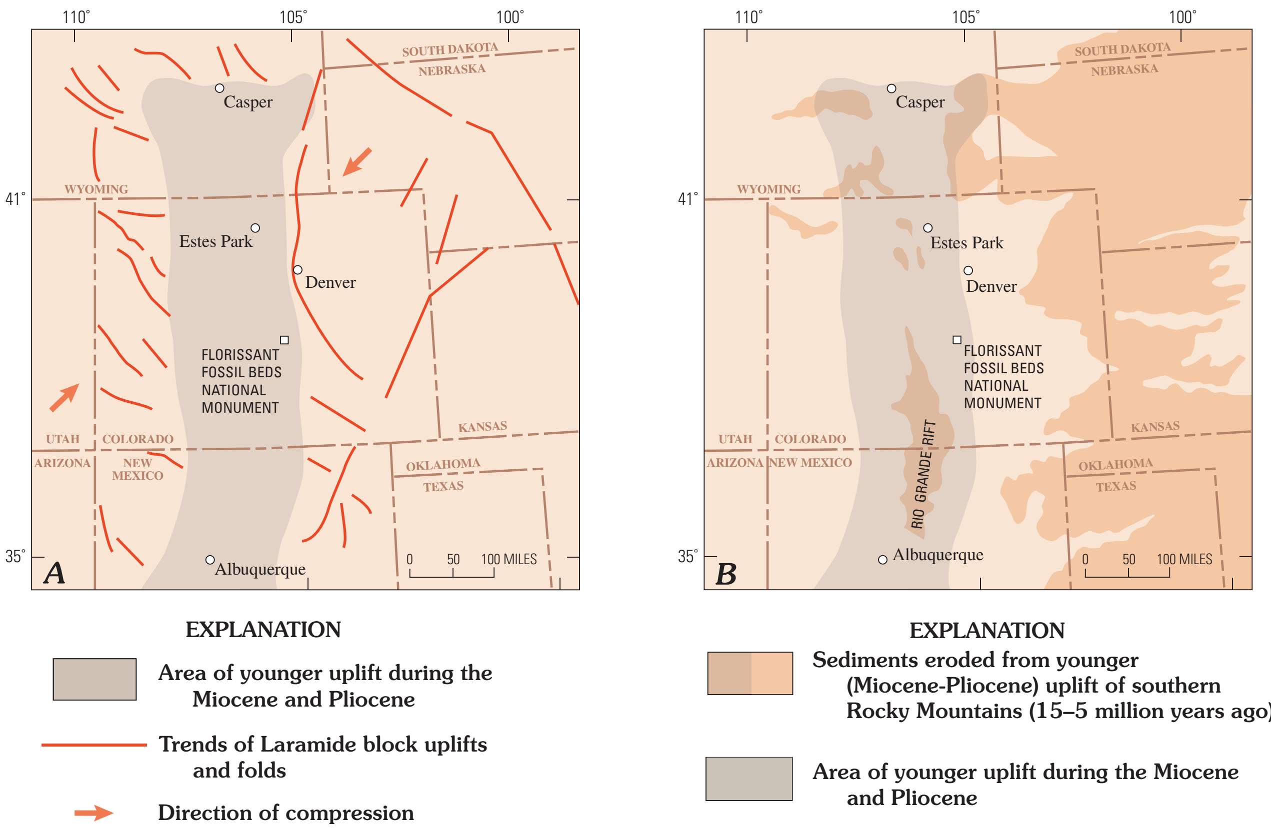

Figure 11. Sketch maps showing structural features of the Neogene Southern Rocky Mountains. *A*, Contrast between trends of Laramide folds and block uplifts versus Neogene uplift of the Southern Rocky Mountains (after Eaton, 1986, 1987; Tikoff and Maxson, 2001). *B*, Distribution of Neogene sediments eroded from the uplifted axis of the Southern Rocky Mountains.

Economic Geology

The geologic resources of the Estes Park quadrangle have afforded wide opportunities for commercial benefit and economic growth over the years. The mountain regions contain deposits of gold, silver, and tungsten that were exploited in the late 1800's and early 1900's. The foothills-hogback zone has been extensively quarried for various stone products over the years. The Denver Basin area has been explored and developed for substantial resources of coal, oil, and natural gas. The valleys of the major rivers and streams have all produced substantial quantities of sand and gravel for building materials, and will continue to do so in the foreseeable future. Above all, water has been the most heavily developed natural resource of the region.

Colorado Mineral Belt

Early Tertiary intrusive rocks emplaced during and following the Laramide orogeny provided heat sources, hydrothermal fluids, and hosts for several significant mineral deposits in the southern part of the Estes Park quadrangle. These mineralized areas constitute the northernmost extent of the Colorado Mineral Belt in this region, as defined and cataloged by Lovering and Goddard (1950). Most of the following summary is based on their report, modified on the basis of new information about rock composition and age (DeWitt, written commun., 2007).

The Boulder County tungsten district (fig. 6) is defined by numerous east-trending veins that cut the Boulder Creek Granodiorite (XgdB) in a narrow zone just north of Middle Boulder Creek several miles west of Boulder (Lovering and Tweto, 1953). The veins were assayed repeatedly for gold and silver in the late 1800's, but they were not developed because they lacked enrichment in these elements. The principal ore mineral is ferberite, the iron-rich variety of the wolframite tungstate series. It forms large, dark, metallic crystals intergrown with quartz, minor scheelite, and sulfides, and the rock adjacent to the veins shows considerable sericitic alteration. Ferberite was finally identified in 1899 after commercial uses of tungsten were identified, and numerous mines were developed throughout the district. Production grew to a peak in World War I due to the need for high-tungsten steels in armament, and continued at a low level through World War II. The tungsten district is not clearly related to Laramide intrusive rocks, but mafic monzonite

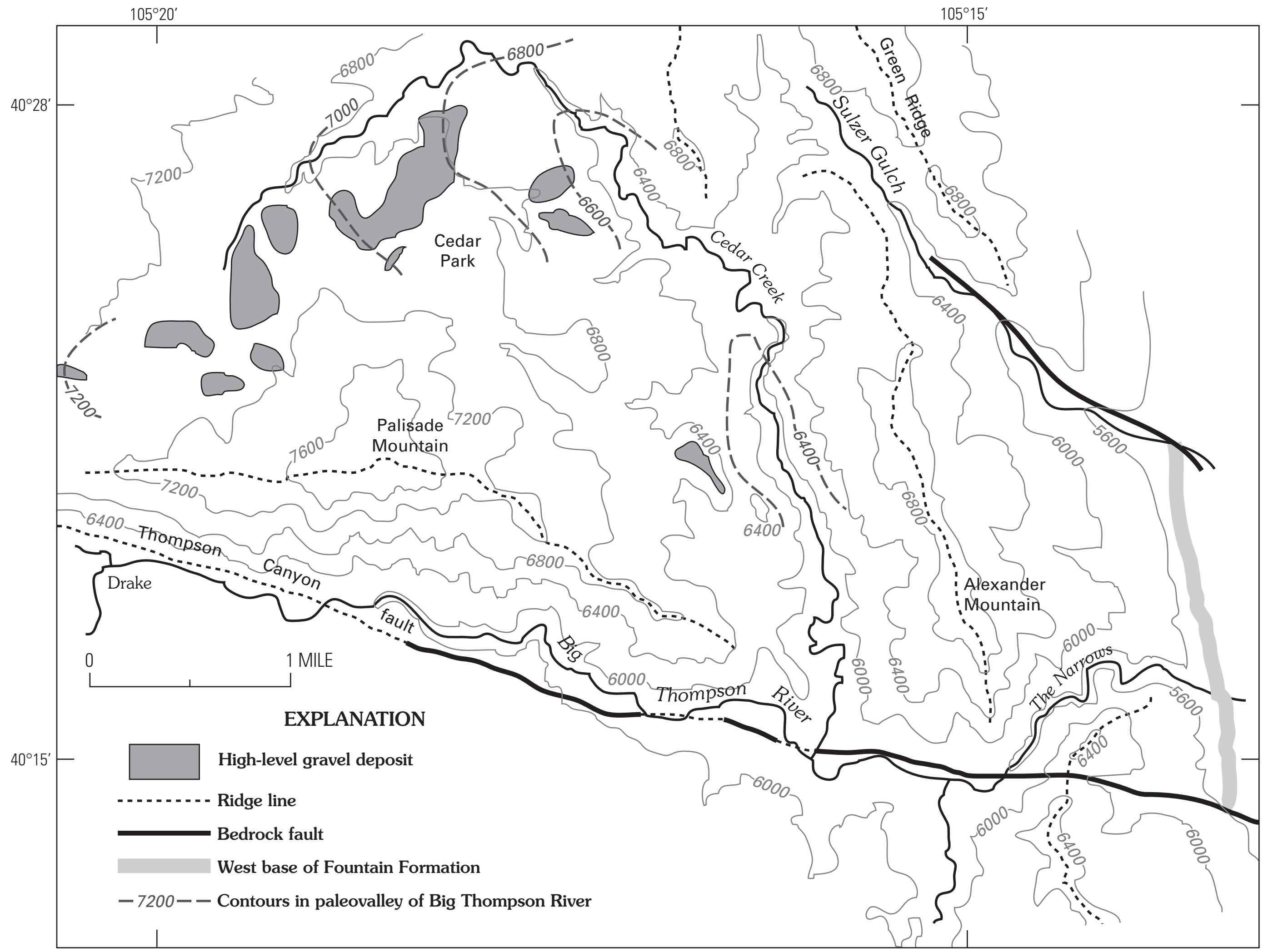

Figure 12. Map showing Miocene(?) high-level gravel deposits and inferred paleodrainage of the Big Thompson River.

porphyries (Late Cretaceous) are spatially associated with this mineralization nearby south of the Estes Park quadrangle.

The other mine districts of the Colorado Mineral Belt (fig. 6) in the southern part of this quadrangle are similar and are clearly related to hydrothermal action surrounding the Laramide intrusive rocks. The principal ores are native gold and silver, precious-metal tellurides, minor arsenides and antimonides, and some base-metal sulfides. The principal gangue minerals are quartz, ankerite, fluorite, and barite. Productive veins are most common in fracture zones and silicified faults within Proterozoic rock adjacent to the Laramide intrusions.

The Ward district was discovered first in 1861, and the nearby Sunset district soon thereafter. Ore shoots occur in fractured basement rock near intrusions of syenite, monzonite porphyry, and quartz latite.

Gold Hill (and surrounding areas) was discovered in 1872 and is known for veins rich in gold telluride minerals, as well as pyritic gold. Ore is preferentially located in and near several large, northwest-trending, silicified breccia zones that traverse the area in proximity to abundant Laramide porphyry dikes.

The Jamestown district is hosted by a large stock of Late Cretaceous monzodiorite (Kmzd) that is, for the most part, only slightly mineralized. Fluids generated by a younger syenite porphyry (₽ɛsy) (about 56 Ma) produced mineralization and zoned patterns of alteration and ore distribution typical of copper-molybdenum porphyry deposits worldwide. Principal value in the district is associated with gold telluride veins and with silver in base-metal sulfides. The Jamestown district also produced significant quantities of fluorite that were deposited in the brecciated collapse zone on the margin of the syenite porphyry.

Coal, Oil, and Gas

The marine deposits of the Cretaceous Western Interior seaway contain abundant carbonaceous debris that served as source rocks for significant oil and gas deposits of the region. Chemically distinctive, but relatively minor, oil is also produced from the Permian Lyons Sandstone (Pl) near Berthoud and Loveland. All oil and gas resources in the Denver Basin are described in detail and summarized by Higley and Cox (2005), the major source for the following summary.

More than 245 million barrels of oil and more than 2.2 trillion cubic feet of natural gas have been produced through 2000 from wells in the Denver Basin since early development began in the late 1800's. The principal source rocks are in the middle part of the South Platte Formation (upper Dakota Group) as well as the shaly formations between the Dakota Group (Kd) and the Niobrara Formation (Kn). Hydrocarbon production is greatest from sandstone bodies within the Dakota Group, at the base of the Niobrara Formation, and within the Upper Cretaceous Pierre Shale (Kp). The high rates of subsidence and sedimentation within the Denver Basin during Late Cretaceous time led to burial, heating, and generation of oil and gas.

Most of the oil and gas production in the Denver Basin comes from the Muddy Sandstone (upper Dakota Group; equivalent to the first sandstone member of the South Platte Formation), also known in the petroleum literature as the "J" sandstone. It is well confined above and below by shale formations (both significant source rocks) and its high permeability enhances production. Traps are a combination of stratigraphic pinch-outs, up-dip stratigraphic truncations, and numerous anticlinal structural traps within the Laramide folds adjacent to the Front Range uplift. Oil-producing wells are concentrated in the southeastern corner of the Estes Park quadrangle in what is known as the Spindle field (mostly in Weld County). The much larger Wattenberg field produces mainly natural gas, and extends far to the east and northeast beyond the quadrangle.

Coal is contained in several economic seams within the Laramie Formation (Kl), a collection of floodplain and swamp deposits that accumulated on the landward side of the regressional Fox Hills Sandstone (Kfh) during the withdrawal of the Late Cretaceous Western Interior seaway. These coals in the lower part of the Laramie Formation were mined at the surface and in the shallow subsurface within the Boulder-Weld coal field as early as the 1850's, and sporadically until the 1970's (Roberts and others, 2001). The following summary is based largely on the comprehensive summary of Roberts (2005).

Approximately 130 million tons of subbituminous-rank coal have been produced from the Denver Basin area during about 120 years of mining. Nearly all of the coal is extracted from the lower 300 ft of the Laramie Formation and is contained in seams that range from 1 to 20 ft thick. Most mines operated from surface excavations down to depths of 400 ft or more (Roberts and others, 2001). The principal area of production within the Estes Park quadrangle is the Boulder-Weld coal field, and it historically produced more than 80 percent of the coal from the entire Denver Basin. Mining ceased in 1979 in this area due to competing uses for the land surface (chiefly housing and commercial centers), and because adequate alternate sources were available in the region. Coal-bed methane has not been produced in this area as of 2007, but increasing demand and market price may lead to exploration in the near future. Estimates of unmined coal and lignite within 3,000 ft of the land surface are substantial.

Sand and Gravel

Quaternary floodplain and fluvial-terrace deposits of sand and gravel have been widely developed within the quadrangle for construction materials (Schwochow and others, 1974). The youngest terraces (Holocene, Broadway Alluvium (Qbr), and some Slocum Alluvium (Qs)) have been most thoroughly developed because the clasts are freshest and post-depositional soils are relatively thin. Sand and gravel quarries are common along the floodplains of the Big Thompson River, St Vrain Creek, and Boulder Creek, as well as along the Colorado and Fraser Rivers west of the Continental Divide.

Quarry Stone

Numerous quarries have been opened in the foothills-hogback belt east of the Front Range to exploit resources of dimension stone and flagstone for building materials. The Lyons Sandstone (Pl) is extensively quarried in and around the town of Lyons because the crossbedded, eolian sandstone is strongly cemented with silica and naturally cleaves into slabs along bedding-plane surfaces.

Sandstone of the Dakota Group (Kd) is also quarried in numerous locations along the range front. It is less dimensional than the Lyons Sandstone, but the Dakota hosts abundant lichen that produce an aesthetically pleasing product referred to as "moss rock."

Sills of Laramide-age dacite and quartz latite crop out in the Paleozoic sedimentary rocks within the foothills-hogback belt between Lyons and Boulder (Wrucke and Wilson, 1967; Braddock, Houston, and others, 1988). One of the largest sills has been extensively quarried southwest of Lyons where it is exposed along the St Vrain Creek. This high-density rock is used chiefly for road base, riprap, and landscape dimension stone.

Water

The earliest settlers in this region recognized that water was a critical commodity to develop and control. For many, the long, slow journey west of the Mississippi River must have provided ample opportunity to observe that average annual rainfall diminished westward and that the native vegetation became progressively more meager closer to the foot of the Front Range. Their first views of the snow-capped peaks of the Continental Divide probably stimulated speculation about means to capture meltwater from the snowpack and hold it in reserve for the summer growing season.

Some of the earliest development projects in this region were ditch systems constructed by settlers during the post-1849 California gold-rush migration. Every major stream that issues from the mountain front is diverted at numerous levels in privately-owned ditches that convey the spring and summer flow to agricultural lands on the Colorado Piedmont. Rights to divert streamflow for beneficial use follow the distinctly western U.S. legal doctrine of prior appropriation, simply expressed as "first in time, first in right". That is, users are afforded shares of streamflow based on priority of use, and the earliest water developers are vested with "senior" water rights in western streams. One of the significant legal cases that confirmed this system in the Rocky Mountains was settled by the 1882 Colorado Supreme Court in a dispute over diversion of the St Vrain Creek (Coffin vs. Left Hand Ditch Company; see Wilkinson, 1992, p. 231–235).

Further water development began in the high mountain valleys during the late 1800's and early 1900's. These projects constructed dams, reservoirs, and conveyances for the purpose of storing winter snowpack so that meltwater could be released during the middle and late summer for irrigation on the piedmont (Cole, 2004a). Many projects included ditches and tunnels in the headwaters of the Colorado River designed to capture and divert Pacific-bound waters to the eastern side of the Continental Divide. The Grand Ditch that traverses the eastern face of the Never Summer Mountains is one of the earliest water-diversion and storage projects that remains in use today. Several small reservoirs were constructed by irrigation companies and piedmont municipalities prior to establishment of Rocky Mountain National Park in 1915. All of these structures have since been removed (Sandbeach Lake, Pear Reservoir, Bluebird Lake, and Lawn Lake) following Park Service acquisition of the water rights. Numerous other small reservoirs remain in the east-flowing drainages of the Front Range to provide water supply to local communities.

By far, the largest trans-mountain water-diversion infrastructure of the region is the Colorado-Big Thompson (C-BT) Project of the Federal Bureau of Reclamation. It was constructed in stages between 1938 and 1957 to meet numerous needs in northern Colorado (Cole, 2004a). It collects and diverts water from the Colorado River headwaters west of the Continental Divide and transmits it through the 13-mile-long Alva B. Adams tunnel beneath Rocky Mountain National Park to a distribution system along the Big Thompson River drainage on the eastern slope. The C-BT Project spans 150 miles of waterways and includes 12 reservoirs, 35 miles of tunnels and siphons, and 6 hydroelectric power plants (Cole, 2004a). The project provides municipal water for 30 communities east of the Continental Divide and irrigation water for approximately 615,000 acres along the Big Thompson, Cache la Poudre, and South Platte River drainages.

The C-BT Project provides reliable water supply, regulated streamflow through seasons and climate changes, environmentally clean hydroelectric power, recreation, additional lakes and canals, agricultural production, economic development, flood control/mitigation, and increased discharges to the South Platte River system and the Missouri-Mississippi River drainage. The trade-offs for this trans-mountain water diversion are decreased flow in the Colorado River headwaters; decreased discharge to the Gulf of California; reductions in wetlands in the upper Colorado River drainage; visual impacts of impoundment structures, power plants, and power lines; evaporation losses from reservoirs; and reduction in peak flood-flows of dammed rivers (Cole, 2004a).

Environmental Geology

Geology is one of the underlying issues that warrants consideration when any part of the landscape is evaluated for a particular use. The geology of the Estes Park quadrangle is both diverse and complex and so the geologic issues that come into play are equally diverse. Different aspects of geology and topography are important in the relatively populated areas of the Colorado Piedmont in contrast to the less populated, but frequently visited mountainous terrain.

This section of the report summarizes some of the geologic processes and conditions that impact human use of the land. It is not intended to be comprehensive, but rather to present geologic factors that have historically led to hazards or damage when they were ignored. Much of the information in this section regarding landslides and expansive soils is summarized from a 2004 field-trip guide covering environmental geology of the Denver metropolitan area (Abbott and Noe, 2004) and from several papers referenced in that guide. The Colorado Geological Survey (CGS) has also published several papers and booklets on geologic hazards that are especially pertinent (Soule and others, 1976; Noe, 1997; Noe and Dobson, 1997; Noe and others, 1997, 1999).

Landslides, Rockfalls, and Debris Flows

Landslides occur in several geologic settings in the Estes Park quadrangle. The large-scale block-glide landslide complexes of the Dakota Group (Kd) strata (Braddock, 1978) were described previously in the section on "Post-Laramide structures;" most are believed to be inactive, although excavation of the toe area could lead to further earth movement. Mountain landslides involving Proterozoic crystalline rocks are known from the Jackstraw Mountain area on the east side of Trail Ridge Road, but they are rare.

Landslides and slumps occur in several areas of the quadrangle in association with rock units that contain abundant silt or expansive clay minerals. The Troublesome Formation (NPEt) in the Granby valley has numerous slumps and slides on moderate slopes that are due to this condition. A large landslide complex involves hydrothermally altered Oligocene intrusive rocks on the northwest slope of Porphyry Peaks, southwest of Grand Lake. The Laramie Formation (Kl) in the southeastern part of the quadrangle is also susceptible to slope failure due to high clay content. The Cretaceous Pierre Shale (Kp) contains several beds of weathered volcanic ash (referred to as bentonite) that are rich in expansive clays that have led to slope failure in the past.

The west side of Boulder is marked by a slow-moving landslide that leads to persistent road-maintenance issues. The westward extension of Baseline Road (southern quadrangle boundary) zig-zags up the face of Flagstaff Mountain and crosses a zone of slumping ground within the strongly tilted sedimentary beds deformed by the local Boulder thrust fault.

Rockfalls are fairly frequent in mountain canyons and occasionally disrupt highways and trails. They are most common during the spring and summer months when rainfall and snowmelt raise the local water table, increase soil pore-water pressure, and allow unstable rock masses to slide and fall.

Debris flows have occurred on several occasions within the quadrangle as a result of elevated pore-water pressure near water-conveyance systems. The historic Grand Ditch was constructed in the 1880's to intercept runoff from the Never Summer Mountains and convey it to the Cache la Poudre drainage for agricultural use in the Greeley-Fort Collins area. This ditch traverses the east slope of the Never Summer Mountains at an elevation of about 10,200 ft and passes through La Poudre Pass and down to Long Draw Reservoir (north of the Estes Park quadrangle). Several debris flows appear to have originated just below the ditch, causing water-charged rock and mud to cascade down the lower slopes of the Never Summer Mountains to the Kawuneechee Valley below. One modern debris flow event occurred in 1978 east of Baker Mountain and impacted the Holzwarth Ranch in the valley below. The deposit of this flow was shown on the geologic map of Rocky Mountain National Park (Braddock and Cole, 1990). A debris flow in the spring of 2005 caused damage to the historic mining camp of Lulu City near the headwaters of the Colorado River, but it originated above the Grand Ditch.

A catastrophic debris flow occurred on the early morning of July 15, 1982, as a result of failure of the Lawn Lake dam at nearly 10,900 ft elevation in the Mummy Range section of Rocky Mountain National Park. The dam had been constructed prior to establishment of the Park in the early 1900's to serve as high-altitude water storage for the City of Loveland and other entities along the eastern Big Thompson River drainage.

Over time, the 80-year-old earthen dam had developed a leak around the outlet pipe, leading to subsurface erosion and ultimately to collapse. When the dam gave way, more than 670 acre-feet of water were released into the Roaring River channel. The torrent cascaded down the 5-mile valley toward Horseshoe Park, eroding a deep 30-ft gash into the channel banks, and picking up boulders, soil, and trees in a fluid slurry. That debris flow reached the mouth of Roaring River in a matter of minutes and deposited a broad alluvial fan at the floor of Horseshoe Park. The maximum flood-stage flow is estimated to have been as high as 18,000 cubic feet per second. The flood transported many boulders more than 3 ft in diameter (as large as 14 × 17 × 21 ft), many of which were eroded from the north lateral glacial moraine of the Fall River drainage.

Expansive Soils

Clay-rich soils in the Colorado Piedmont section of the Estes Park quadrangle have caused damage to foundations, structures, and roads due to expansion and contraction resulting from alternate wetting and drying. Flat-lying and gently inclined beds of the Pierre Shale (Kp) and the Laramie Formation (Kl) are most susceptible to this process due to the presence of zones rich in expansive clays. Experience during the last 40 years has shown that these conditions can be mitigated with engineered systems in most cases, and all county building departments require soil testing and appropriate mitigation for new construction in areas east of the foothills-hogback belt.

Floods

The mountainous terrain of the Front Range is drained by major rivers that flow through deeply incised, narrow canyons between about 9,000 ft and 5,500 ft elevation. Major summer thunderstorms concentrate their most intense rainfall between elevations of about 7,000 ft and 9,000 ft due to orographic conditions, and so the potential for mountain-valley flash flooding is both high and persistent (Cole, 2004a). Most of the incised canyons in this region contain no flood-control dam below the high-rainfall zone, largely because most of the deep canyons are major transportation routes and have been since settlement began in the 1800's (Button Rock Reservoir on North St Vrain Creek is the sole exception). Every major stream on the east slope of the Front Range has experienced moderate to significant flooding at one time or another in recorded history, and the likelihood of future flooding is high. Modern development has encroached further on floodplains and floodways, and so the potential for damage continues to rise.

The Big Thompson flood of 1976 is instructive in this regard (see hydrologic and geologic summaries; McCain and others, 1979; Shroba and others, 1979). The summer afternoon of July 31 began with typical thunderstorm activity that centered around the head of the deep-valley section of the Big Thompson River below the dam at Lake Estes. The thunderstorm grew in size and remained stationary in this position, resulting in more than eight inches of rainfall over large portions of the drainage system within a period of little more than one hour (McCain and others, 1979; Cole, 2004a). The resulting flash flood swept rapidly through the Big Thompson River canyon, destroying numerous buildings, bridges, roadways, and other structures, and drowning more than 143 people. Most of the land downstream from the mouth of the canyon was agricultural and sparsely settled and so damage and loss of life were rare east of the mountain front.

The conditions are quite different for Boulder Creek and St Vrain Creek. The communities of Boulder and Lyons, respectively, have grown up around the canyon mouths of these major streams and many structures lie within the geologically defined flood plains. The last major flood in Boulder occurred in the summer of 1894, during which large portions of low-lying sections of the downtown were flooded and several creek-side structures were lost. Lyons has also been spared major flooding for decades, but the potential for a 1976-Big Thompson flood event is significant over decadal intervals.

Failure of the Lawn Lake dam in 1982 (described above in the discussion of debris flows) produced flooding downstream from the alluvial fan deposition site and through the town of Estes Park. As the flood discharge spread out across the low-gradient Horseshoe Park, it deposited significant sand and silt more than 0.5 mile downstream (Blair, 1987). At the eastern end of Horseshoe Park, the flow returned to a narrow channel cut in bedrock and cascaded down toward the town of Estes Park, picking up more debris in the process. Thanks to advance warning, the downtown area was evacuated before the flood surge arrived. No lives were lost in town, but 18 bridges were destroyed and most of Estes Park was buried in 2 ft of mud and debris. The flood discharge was contained in the Lake Estes reservoir.

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